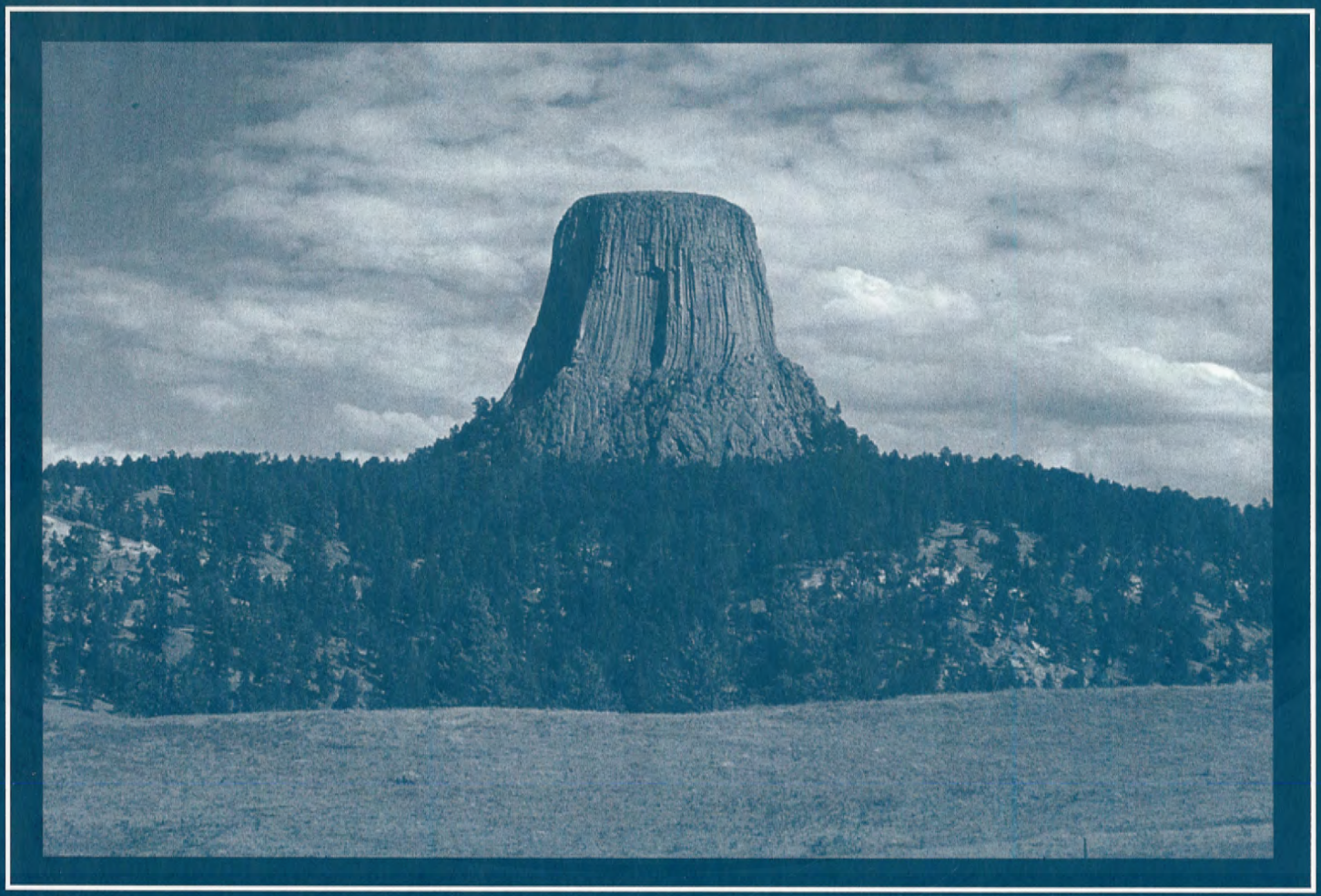
WYOMING STATE GEOLOGICAL SURVEY
Gary B. Glass, State Geologist

Proceedings of the 32nd Annual Forum on the Geology of Industrial Minerals

edited by

Richard W. Jones and Ray E. Harris



Public Information Circular No. 38

1997
Laramie, Wyoming

Annual Meetings of the Forum on the Geology of Industrial Minerals

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17th	1981	Albuquerque, New Mexico
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19th	1983	Toronto, Ontario, Canada
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26th	1990	Charlottesville, Virginia
27th	1991	Banff, Alberta, Canada
28th	1992	Martinsburg, West Virginia
29th	1993	Long Beach, California
30th	1994	Fredericton, New Brunswick/ Halifax, Nova Scotia, Canada
31st	1995	El Paso, Texas
32nd	**1996**	**Laramie, Wyoming**
33rd	1997	Quebec City, Quebec, Canada
34th	1998	Norman, Oklahoma
35th	1999	Salt Lake City, Utah
36th	2000	England, United Kingdom

Front Cover: Devils Tower is a spectacular erosional remnant of a Tertiary volcanic neck on the western edge of the Black Hills uplift in northeastern Wyoming. The tower, which is composed of nepheline syenite (phonolite), has intruded relatively flat-lying Mesozoic sedimentary rocks (foreground and near base of tower). Field trip #4 of the 32nd Annual Industrial Minerals Forum briefly (between the rainstorms) visited Devils Tower. For further description of Devils Tower and associated Laramide-age magmatism, see article by Snoke, this volume, p. 26 and 28. Photograph courtesy of Wyoming Department of Transportation.

WYOMING STATE GEOLOGICAL SURVEY
Gary B. Glass, State Geologist

Public Information Circular No. 38

PROCEEDINGS OF THE 32ND ANNUAL FORUM ON THE GEOLOGY OF INDUSTRIAL MINERALS

Meeting in Laramie, Wyoming, May 19-21, 1996
With field trips on May 19 and 22-25, 1996

edited by

Richard W. Jones and Ray E. Harris

Laramie, Wyoming

1997

WYOMING STATE GEOLOGICAL SURVEY

Gary B. Glass, *State Geologist*

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Preface

The Proceedings Volume of the 32nd Annual Forum on the Geology of Industrial Minerals is dedicated to the memory of Marion Bates, wife for 58 years of Industrial Minerals Forum founder Bob Bates and herself an enthusiastic supporter of the Annual Forum on the Geology of Industrial Minerals. She was a familiar participant in the Forum and at guest events. Mrs. Bates died in Columbus, Ohio, September 26, 1995.

The 32nd Annual Forum on the Geology of Industrial Minerals was held at the University of Wyoming in Laramie, Wyoming, May 19-21, 1996. The event was hosted by the Wyoming State Geological Survey (WSGS), with assistance from the University of Wyoming Office of Conferences and Institutes. The geology and industrial minerals of Wyoming were featured, although regional developments, including presentations of the changing industrial mineral industries of the states that border Wyoming were also given. Technical presentations of national and international interest, including new products, new technologies, and environmental issues were an important part of the program.

The 32nd Forum was attended by 204 participants, including guest speakers and poster presenters. These came from 33 U.S. states, 5 Canadian provinces, and 3 other countries. The 32nd Forum can be said to be the highest Forum yet, since it was held at Laramie's 7200-foot elevation.

Forum events

The Forum icebreaker was held in the Crane-Hill Dining Hall on the University of Wyoming campus. It offered a chance to meet past Forum acquaintances and introduce first time attendees to the warmth of the Forum. Jazz musicians from the Laramie Junior High School Jazz Band, Dave Rickard conducting, provided entertainment.

Introductions and welcoming addresses were given prior to the technical sessions by State Geologist Gary Glass for the State of Wyoming and by Jay Fromkin, Director of University Relations, for the University of Wyoming. Dick Berg, Peggy Dahlheim, and Roger Sharpe were recognized as the Steering Committee for the 32nd Forum, and George Austin gave a remembrance for Bob and Marion Bates.

Technical sessions were held in the Grand Ballroom in the Wyoming Student Union on the University of Wyoming Campus. Sessions were divided into: Keynote addresses: The geology of Wyoming, chaired by Dick Jones of the WSGS; Industrial minerals of Wyoming, chaired by Ray Harris of the WSGS; Industrial minerals of the Rocky Mountains and High Plains, chaired by Arnold Doden, Bob Bates Scholarship recipient; Land use topics, also chaired by Arnold Doden; and Environmental and other topics, chaired by Arwen Harris of the University of Pittsburgh.

Some participants elected to attend a western barbecue Monday night at the historic Overland Ranch, seven miles west of Laramie. Despite the freezing May evening in the Laramie Valley, those in attendance enjoyed good food, Wyoming history, and the old west atmosphere. Diners sat on hay bales to eat and heard a talk by Marge Richardson, owner of the Overland Ranch, on the history of the ranch and the surrounding area; State Geologist Gary Glass then discussed the registration of geologists in Wyoming and other states.

The closing banquet was held in Crane-Hill Dining Hall. In addition to introducing past and future Forum hosts and the recipient of the Bob Bates Scholarship, Arnold Doden, diners were treated to singing by the Laramie High School Plainsmen Singers, Sandy Steele directing, and an entertaining talk on Wyoming dinosaurs by the Director of the University of Wyoming Geological Museum, Brent Briethaupt.

Field trips

As is true of most Forums, field trips were an important part of the meeting. Approximately 90% of the 32nd Annual Forum registrants participated in at least one field trip.

Due to a last minute cancellation of bus transportation, University of Wyoming and State of Wyoming vans, suburbans, and station wagons carried 60 participants on the pre-meeting field trip to the Sloan kimberlite in northern Colorado, south of Laramie. Field trip leader Dan Hausel of the WSGS discussed the emplacement, discovery, and potential of diamonds in the State Line district and other areas in the United States.

Following the forum, three field trips were scheduled. A bus load of participants (43) visited five industrial mineral producing and fabricating localities in southeastern Wyoming, including the state's only cement plant, the region's largest commercial aggregate quarry (see photograph, page 106), a pioneering granite fabricating plant, and Georgia Marble's white marble quarry and processing plant. Field trip leader

Steve Monell of Colorado Lien smoothed the way for the trip. Participants also enjoyed a lunch of regionally famous Chugwater Chili at chili headquarters in Chugwater, Wyoming.

Twenty-four participants visited the Wyoming trona patch and toured underground trona mines and surface refining facilities (**Figure 1**). Field trip leader R. Terry Leigh of Tg Soda Ash arranged tours at three of the five producing localities in the trona patch. Guests stayed at Little America, an isolated motel and restaurant complex in the middle of the sagebrush-covered prairie typical of the trona mining area.

Twenty-seven participants (**Figure 2**) braved four days of thunderstorms, rain, mud, fog, and, yes, snow, to visit industrial mineral producing localities in eastern and northern Wyoming. Despite the weather, all sites in the itinerary were visited, although views from many scenic overlooks were of clouds or falling rain, fog, or snow. Localities visited included Guernsey Stone's dolomite aggregate quarry near Hartville (see photo, page 166); Pacer Corporation's feldspar processing plant in Custer, South Dakota; American Colloid's Colony East, Wyoming, bentonite processing plant; Cyprus-AMAX's Eagle Butte surface coal mine near Gillette, Wyoming, where participants were treated to a view of a 140-foot-thick coal seam and associated clinker aggregate quarry; and a half day at Georgia Pacific's Gypsum quarry and wallboard plant south of Lovell, in the Bighorn Basin of northwestern Wyoming.

Numerous sites of industrial mineral and geologic interest were visited, including the limestone sequence in Wind River Canyon, the site of the photograph in **Figure 2**. Also visited were points of interest including Mount Rushmore in a thunderstorm, Devils Tower (**Cover** photograph), where the fog miraculously lifted for the hour we were there, the Bighorn Mountains in snow, and Thermopolis Hot Springs in a downpour. Pronghorn antelope, deer, moose, and even Rocky Mountain goats were occasionally seen.

Overnight stays at luxury motels in Spearfish, South Dakota; Sheridan, Wyoming; and Thermopolis Hot Springs, Wyoming, rested and refreshed the tired participants. Some participants dined at the historic Sheridan Inn, and a wrap-up wine and cheese party was provided by the Thermopolis Hot Springs Holiday Inn, where guests amused themselves by trying to identify the collection of worldwide trophy game animals in the lounge. Percy Hastings, from Bulawayo, Zimbabwe, somehow was able to identify more than anyone else. Tom Westervelt's experience when the roof of his room at Hot Springs State Park collapsed in the middle of the night due to the heavy rain is worth mentioning. The motel responded appropriately by moving Tom into the only unoccupied room,



Figure 1. FMC Wyoming Corporation's Westvaco trona refinery and unit train load out facility, Green River Basin, Wyoming (photograph courtesy of FMC Wyoming Corporation).



Figure 2. Participants of Field Trip #4, 32nd Annual Forum on the Geology of Industrial Minerals, in Wind River Canyon, Wyoming, May 25, 1996. Participants pictured are (left to right): Barrett Williams, Dick Jones, Percy Hastings, Claudette Simard, Tony Roberts, Isadore Million, Tom Westervelt, James McCann, Jim Barker (kneeling), Gretchen Hoffman, George Austin, Marjorie Austin, Otis Clarke (kneeling), Henri-Louis Jacob, John McGee, Tom Hiscox, Arwen Harris, John Alfors, Pat Alfors, Mary Fox, Ray Harris, Cameron Fox. Not pictured: Alex Glover, who took the photograph.

which was, however, the Honeymoon Suite. Tired participants headed home or spent Saturday night in Laramie, while the trip drivers, Arwen Harris, Ray Harris, and Dick Jones, returned the extremely muddy vans to the University Motor Pool.

Guest trips have become important as more Forum participants bring family or friends. Guest Program Chair Mary Harris led guests on tours of museums, shops, and places of historical interest in the Laramie area. Of special interest was a bull whip demonstration, indoors in the University of Wyoming Centennial Complex, by the niece of Calamity Jane of Western Frontier fame. The Centennial Complex is the home of the International Archive of Economic Geology, where all archives for the Forum on the Geology of Industrial Minerals are stored. Also visited was the Wyoming Territorial Park, site of the restored Wyoming Territorial Prison and a tent city, "Hell on Wheels," from the days when the Union Pacific Railroad was constructed through Laramie and the law was the six-gun.

Fees collected but not used in the course of the meetings, field trips, or proceedings volume were given to the Bob Bates Scholarship Fund, established to provide funds for a student(s) to attend a Forum. The amount donated from the 32nd Annual Forum was $7546.69. Most of this money came from the last minute cancellation of field trip buses, forcing the use of University of Wyoming Motor Pool vans, but these were less costly than the buses originally contracted.

Proceedings volume

The papers in this volume were lightly edited by WSGS Editor Richard W. Jones and 32nd Forum Chair Ray E. Harris. Appreciation is given to WSGS personnel Janet A. Van Nuys, WSGS Editorial Assistant, for word processing and layout, and Phyllis A. Ranz and Fred H. Porter, III, Cartographers, for computer drafting and cartography.

Finally, a special thanks is given to all of the speakers who took the time to prepare the articles presented in this volume. The articles were very well written and reflect a high degree of professionalism. We appreciate your efforts and hope you are satisfied with this result.

Ray E. Harris,
General Chair and Proceedings Volume co-editor.

Richard W. Jones,
Technical Chair and Proceedings Volume co-editor.

Acknowledgments

The 32nd Annual Forum on the Geology of Industrial Minerals was sponsored and supported by the Wyoming State Geological Survey, including Wyoming State Geologist Gary B. Glass. Financial support in the form of grant funding from the Society of Economic Geologists Foundation, Inc., and The Albany County (Wyoming) Board of Tourism is acknowledged and appreciated. Additional financial support from the Laramie Economic Development Corporation, Longpre Distributing, the Mineral Lab, Inc., and the Mountain Cement Corporation is also acknowledged and appreciated. Registration, logistics, and the Forum budget were ably coordinated by the University of Wyoming Office of Conferences and Institutes and we appreciate the efforts of Wil Radosevich and Barbara Barnes and their staff.

Companies throughout the region agreed to provide tours of their operations for the 32nd Forum Field Trips. The enthusiastic support of these hosts, listed alphabetically below, is appreciated.

Special thanks are due to the members of the 32nd Forum Steering Committee: Richard Berg, Peggy Dahlheim, and Roger Sharpe, for their advice and assistance in planning and supporting this Forum. Non-voting members of the Steering Committee who also provided advice and assistance include George S. Austin, A. James McCann, Ken Johnson, and Bryce Tripp. Field trip leaders as noted in the Preface included Arwen Harris, Ray E. Harris, W. Dan Hausel, Richard W. Jones, R. Terry Leigh, and Steve Monell. Richard W. Jones also deserves special mention for his time and effort in ensuring the quality and timely publication of this proceedings volume.

American Colloid – Colony East, Wyoming, bentonite plant and mine

Cominco – Sloan diamond property, State Line district, Colorado-Wyoming

Cyprus-AMAX – Eagle Butte coal mine and clinker pit, Gillette, Wyoming

FMC – Westvaco trona mine and processing plant, Green River, Wyoming

General Chemical – Green River, Wyoming, trona mine and plant

Georgia Marble – Wheatland, Wyoming, white marble aggregate quarry and plant

Georgia-Pacific Gypsum Co. – Lovell, Wyoming, gypsum mine and wallboard plant

Guernsey Stone – Guernsey, Wyoming, aggregate quarry

Meridian Aggregates – Granite Cañon, Wyoming, aggregate quarry

Mountain Cement – Laramie, Wyoming, cement plant

OCi – Big Island trona mine and processing plant, Green River, Wyoming

Pacer Corp. – Custer, South Dakota, feldspar processing plant

Sunrise Stone – Wheatland, Wyoming, dimensional stone plant

Geologic history of Wyoming within the tectonic framework of the North American Cordillera[1]

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Abstract

Wyoming, a mountainous highland, is situated along the eastern margin of the U.S. Cordillera. A portion of the northern Great Plains forms the eastern part of the state, and much of the Wyoming Basin and parts of the Southern and Central Rocky Mountains comprise the rest of the state. The early geologic history of Wyoming involved complex accretion and subsequent rifting of an Archean craton, commonly referred to as the Wyoming province. Late in the Early Proterozoic, an oceanic-arc terrane was accreted to the southern margin of the Wyoming province during an oblique collision. The Proterozoic suture, called the Cheyenne belt for exposures in southeastern Wyoming, is a zone of penetrative deformation. Middle Proterozoic anorthositic and associated monzosyenitic to granitic rocks intruded this tectonic zone.

After the mid-Proterozoic, a nearly billion-year gap exists in the geologic record of Wyoming. Lower and middle Paleozoic shelf-facies strata, commonly separated by disconformities, form a veneer upon Precambrian basement rocks. Uplifts of the late Paleozoic Ancestral Rocky Mountains extended northward into southeastern Wyoming and locally were eroded to Precambrian rocks; Pennsylvanian rocks therefore lie directly on the basement in parts of these uplifts. Late Paleozoic and early Mesozoic strata record diverse paleoenvironments, including paralic, eolian, and fluvial settings. Beginning in the mid-Cretaceous, a persistent interior seaway extended from the Arctic Ocean to the Gulf of Mexico. A nearly coeval, thin-skinned, fold-and-thrust belt (Sevier orogenic belt) evolved west of this seaway and remained active into the early Eocene. East of the fold-and-thrust belt, during the Late Cretaceous through early Eocene, deep-rooted reverse and thrust faults fractured the craton, forming basement-cored uplifts separated by deep, actively subsiding basins. These juxtaposed structures are the classic features of the Laramide orogeny of the eastern Rocky Mountains.

Locally important magmatism, including both alkaline and subalkaline rocks, formed volcanic piles and subvolcanic intrusive complexes during the middle to late Eocene. The Heart Mountain detachment fault evolved during the middle Eocene, and allochthonous Paleozoic sedimentary and Tertiary volcanic rocks spread southeastward into the Bighorn Basin from a steep breakaway fault along the northeastern flank of the northern Absaroka Range. Extensive lakes also developed during the middle Eocene in southwestern, central, and northern Wyoming. During the latest Eocene through middle Miocene, airborne volcaniclastic debris, chiefly derived from the Great Basin, fell over the entire Wyoming foreland. The aggradational cycle continued intermittently throughout much of the Neogene, gradually burying all but the highest parts of the Laramide tectonic landscape.

[1]This summary is a revised and updated version of a paper originally published under the same title (Snoke, 1993) in *Geology of Wyoming*, Wyoming State Geological Survey Memoir 5, p. 2-56.

Although late Neogene normal faulting occurred throughout Wyoming, only modest extension resulted. During the last 5 million years, the uplifts have been exhumed, exposing the Laramide skeleton as the present mountainous topography. Debate focuses on the relative importance of the roles of tectonic uplift versus climatic change in this exhumation. High-level erosional surfaces are common in the Rocky Mountains. Most of the high-level surfaces in Wyoming formed late in the Cenozoic, as erosional remnants of previously much more extensive pediments.

The massive Yellowstone Plateau is the product of dominantly rhyolitic volcanic eruptions that occurred during the past 2.2 million years associated with uplift above an inferred mantle hotspot. Recent seismic activity is common near the western border of Wyoming (Intermountain seismic belt); historic seismic events have been comparatively infrequent elsewhere in the state.

Introduction

A remarkably diverse and complex geologic history is preserved in the parallel-sided political unit called Wyoming. Its geologic history is vividly portrayed in color and form on the *Geologic map of Wyoming* by Love and Christiansen (1985). Wyoming provides rich slices of Earth history, preserved as the various rock units and structures that have formed the subjects of countless geological studies. An objective of this article is to put this long, multi-stage geologic history into a continental perspective by integrating the geology of Wyoming into the tectonic evolution of the North American Cordillera (**Figure 1**). Such a complex and broad-ranging task, however, is destined to be an incomplete analysis.

The State of Wyoming includes a large part of the Central Rocky Mountains (or Middle Rocky Mountains; Fenneman, 1931), much of the Wyoming Basin province, several prongs of the Southern Rocky Mountains, and a part of the northern Great Plains. Wyoming is therefore situated in an important Phanerozoic tectonic transition zone, extending from the essentially flat-lying rocks of the continental interior to the folded and faulted strata of the Rocky Mountains. However, this history is superposed on a staggeringly long Precambrian history that can be traced back to remnants of some of the oldest known rocks in North America (Aleinikoff and others, 1989; Mueller and others, 1992; Frost and Frost, 1993).

The present physiography of Wyoming provides important clues to past events that shaped the geologic evolution of the state. The mean elevation of Wyoming is high (about 6700 feet), second only to Colorado among the fifty states (Larson, 1978). The highest elevations in Wyoming are in the Wind River Range, where several peaks are over 13,000 feet; Gannett Peak, the highest in the state, reaches 13,804 feet. The lowest elevation, 3125 feet, is along the Wyoming-South Dakota border. Much of Wyoming is characterized by broad intermontane basins surrounded by massive uplifts with Precambrian rocks exposed in their cores. Along the western margin of the state, Precambrian rocks are not exposed at the surface, and long linear color bands on the geologic map chiefly represent Paleozoic and Mesozoic rocks imbricated by thrust faults. In the northwestern corner of the state, a great plateau of Cenozoic volcanic rocks is the prominent physiographic feature, whereas in northeastern Wyoming the margin of a regional domal uplift, centered in western South Dakota, is the dominant large-scale geologic feature.

Wyoming therefore can be subdivided into several distinct geologic provinces, which include: the fold-and-thrust belt, the uplifts and basins of the Rocky Mountain foreland, the Yellowstone Plateau, the Absaroka volcanic field, the Black Hills, and the Denver-Julesburg Basin (**Figure 2**). My approach to this complex geologic framework is to develop, chronologically, the main tectonic events preserved in the rock record (**Figure 3**).

The Wyoming province: Accretion of an Archean microcontinent

In Wyoming, Archean rocks are extensively exposed in the Laramide basement-cored uplifts of the Rocky Mountain foreland (**Figure 2**). These uplifts provide an important window into the Archean crust that underlies much of Wyoming, a terrane appropriately referred to as the Wyoming province (Engel, 1963; Condie, 1969). The Wyoming province is flanked by Proterozoic orogens on its northern, eastern, and southern margins (**Figure 4**). Its western margin, which extends into northeastern Nevada

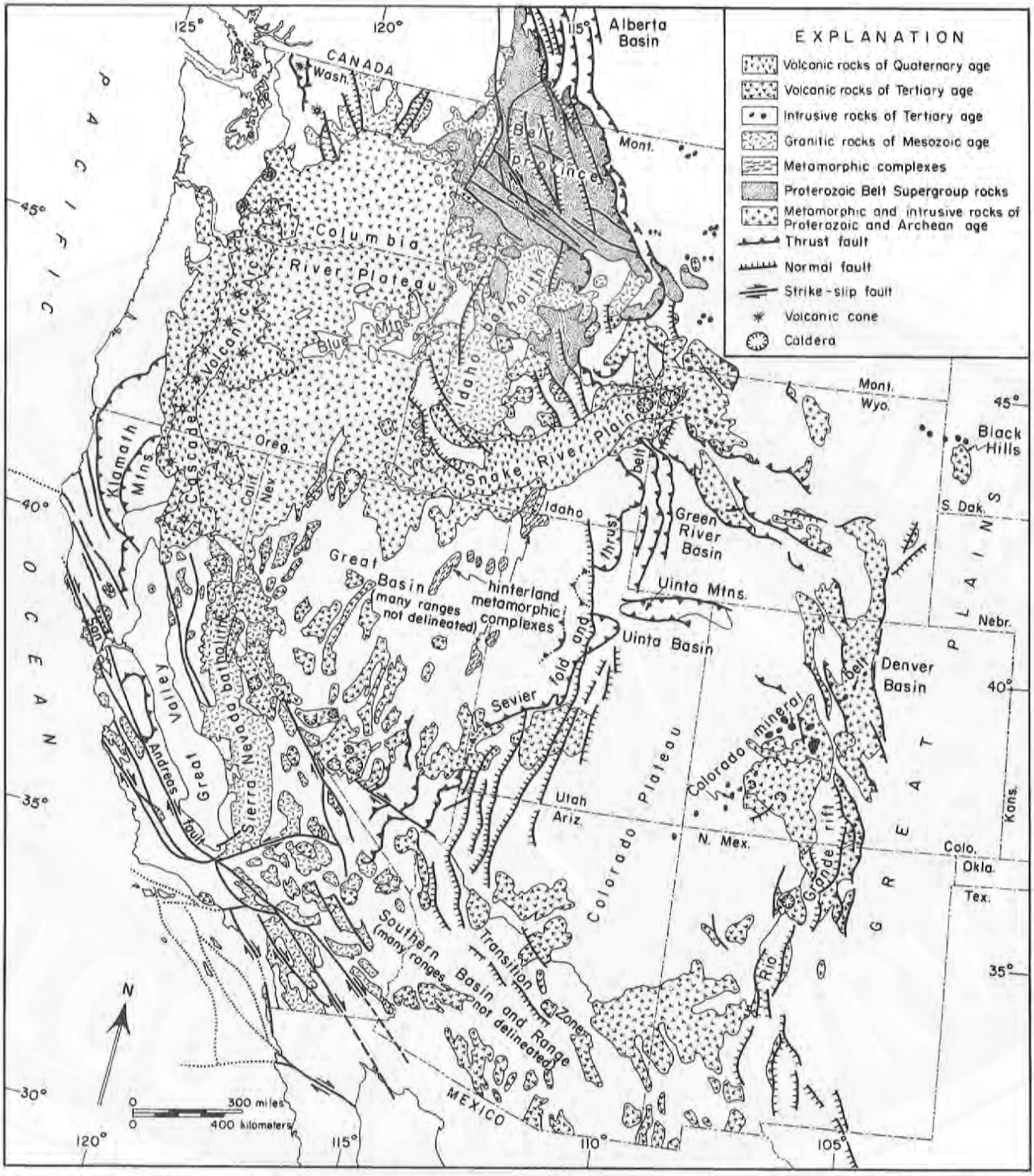


Figure 1. Generalized map of selected tectonic and geologic features of the conterminous U.S. showing Wyoming's location within this regional setting (adapted from King, [1967] 1970).

(Lush and others, 1988) and under the Snake River Plain (Leeman and others, 1985), formed in the late Precambrian during breakup of the Proterozoic supercontinent (Stewart, 1976; Hoffman, 1991).

The Wyoming province is a complex aggregate of various Archean components chiefly including: Early(?) Archean to Middle Archean migmatitic gneisses, commonly containing enclaves of older supracrustal rocks; Late Archean supracrustal sequences; widespread Late Archean granitic rocks (**Figure 5**); and various ultramafic to mafic intrusive rocks. Zircons from a granulite-facies, migmatitic gneiss ex-

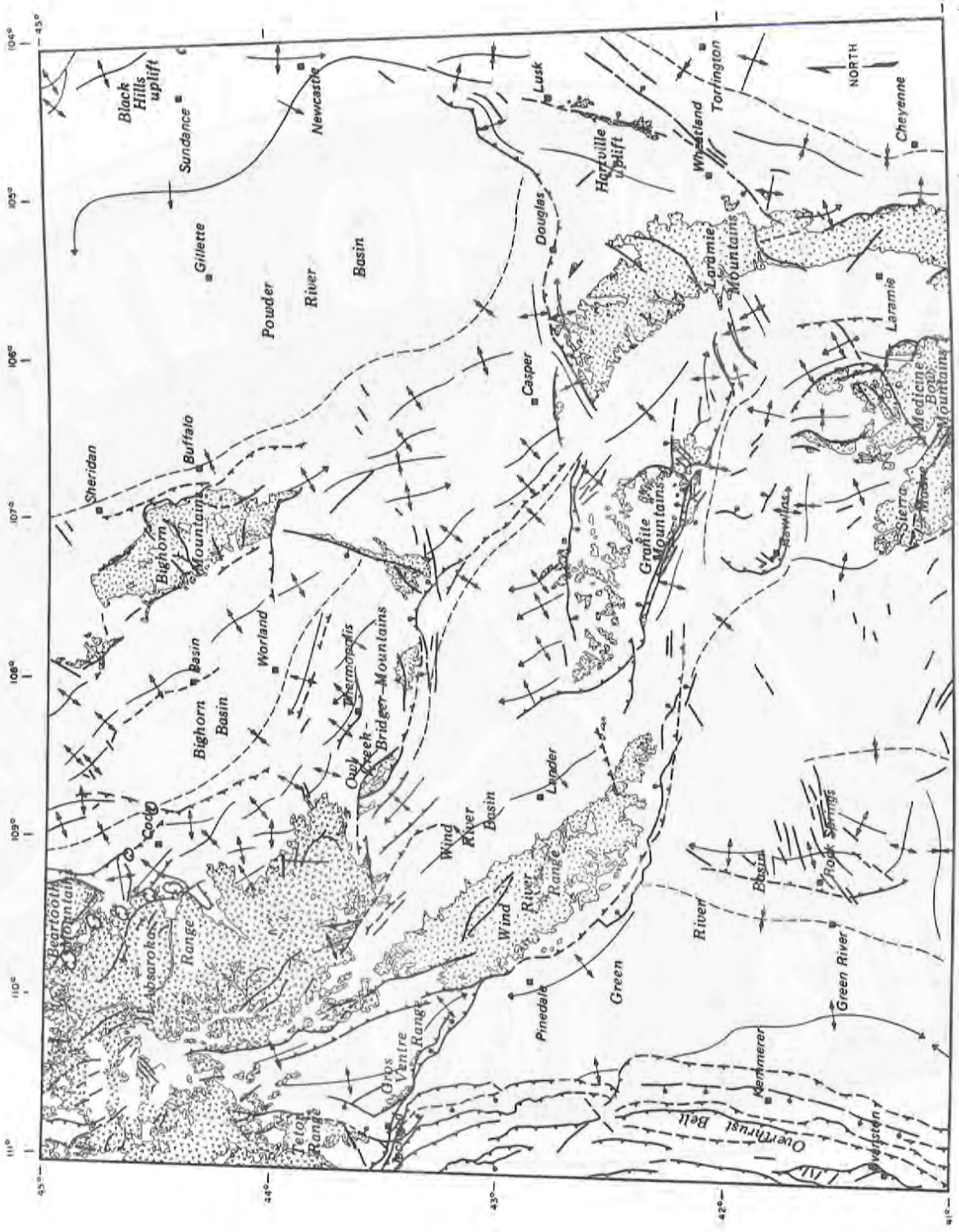


Figure 2. Generalized structural map of Wyoming, showing positions of basement-cored uplifts and main structural features (from Blackstone in Roberts, 1989).

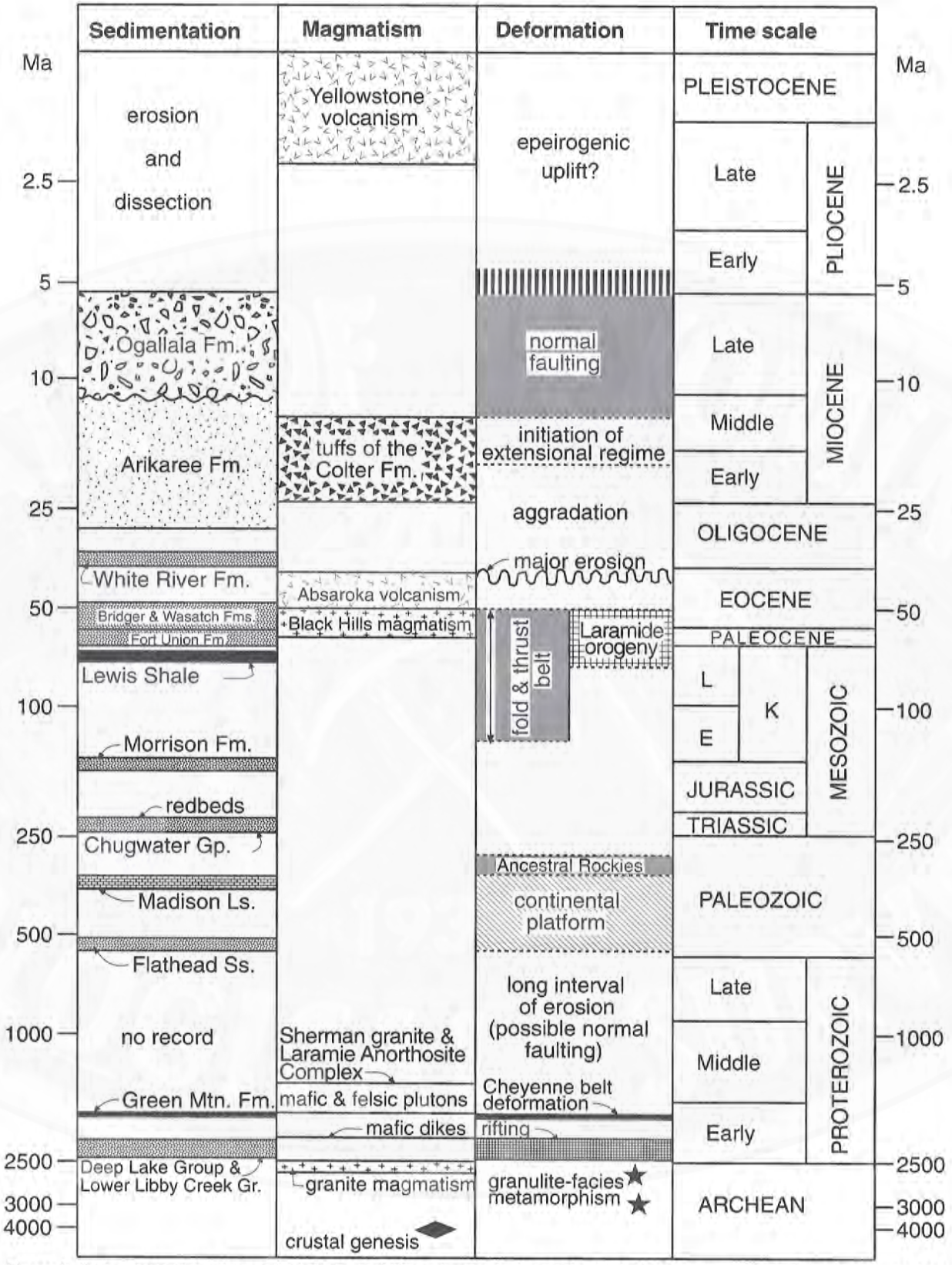


Figure 3. Salient aspects of the geologic history of Wyoming, including sedimentation, magmatism, and deformation. Only selected stratigraphic units are delineated in the sedimentation column. Logarithmic time scale in Ma (millions of years ago); geologic time scale after Palmer (1983). This diagram was designed after Dickinson (1991, figure 5). The age range of Late Archean granitic rocks is derived from Peterman and Hildreth (1978) and Stuckless and others (1985). The age range of the Colter Formation and initiation of normal faulting is from Barnosky (1984). Other dates are derived from various publications cited in the text.

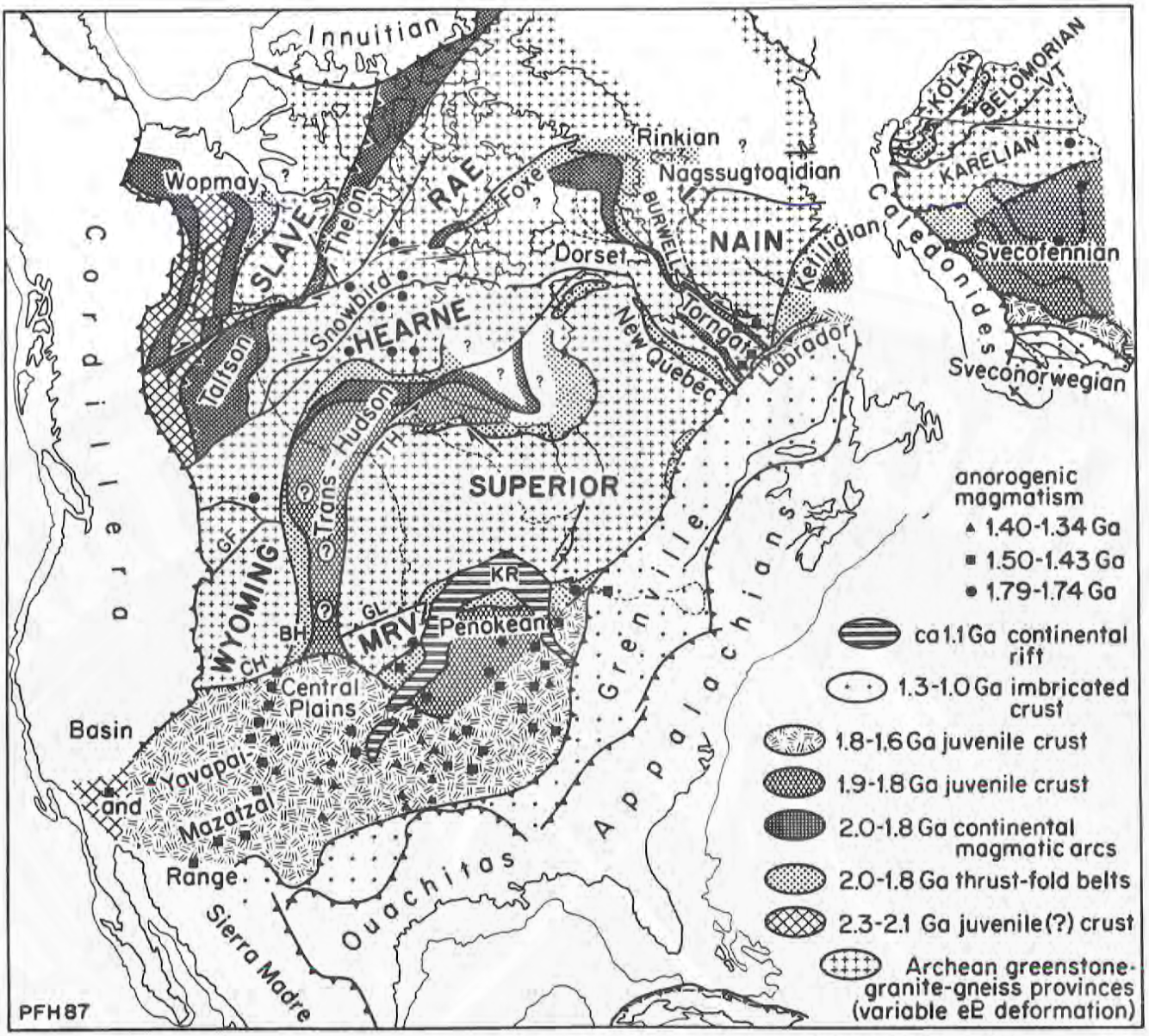


Figure 4. Position of the Archean Wyoming province relative to other Precambrian-age belts and 1.4 to 1.5 Ga magmatism [modified from Hoffman (1989b), figure 13; used with permission]. The platform cover of the North American craton has been removed. Upper-case names are Archean provinces; lower-case names are Proterozoic and Phanerozoic orogens. BH, Black Hills inlier; CH, Cheyenne belt; GF, Great Falls tectonic zone; GL, Great Lakes tectonic zone; KR, Keweenawan rift; MRV, Minnesota foreland.

posed in the central Wind River Range have yielded an Early Archean age component of >3.6 Ga (billion years old) (Aleinikoff and others, 1989). The supracrustal enclaves, a common component of the migmatitic gneisses, therefore probably constitute the oldest known element of Earth history preserved in Wyoming. This dismembered ancient stratigraphy includes metamorphosed igneous rocks (metabasite and metaperidotite) as well as a diverse original sedimentary succession now metamorphosed to pelitic paragneiss, iron formation, and scarce calc-silicate paragneiss and quartzite (Koesterer and others, 1987; Wooden and others, 1988). Metamorphic and geochronometric studies suggest an early history of granulite-facies metamorphism for some of these rocks prior to incorporation into the migmatitic gneiss. The ages of most of the migmatitic gneisses are uncertain or poorly constrained, but most geochronometric studies yield dates within the interval 2.8 to 3.1 Ga (Frost and Frost, 1993). In the Beartooth Mountains, Montana, zircons from a Middle Archean quartzite indicate that an older crustal component (about 3.96 Ga) also exists or did exist in the northern Wyoming province (Mueller and others, 1992).

Late Archean supracrustal sequences are preserved locally in the Wyoming province. Some of these successions are metavolcanic rich and exhibit characteris-

tics of classic greenstone belts of other Archean terranes (e.g., Superior province of the Canadian shield or Kaapvaal craton of South Africa). However, the limited areal distribution and generally higher metamorphic grade of Wyoming province rocks are major dissimilarities. Nevertheless, some mafic metavolcanic rocks in Wyoming's supracrustal sequences are magnesian, with geochemical characteristics of komatiitic basaltic rocks (Snyder and others, 1990). Spinifex-textured mafic metavolcanic rocks have been reported from the Seminoe Mountains in south-central Wyoming (Klein, 1980; Snyder and others, 1989). Still other Late Archean successions are characterized by stromatolitic metadolomite, mafic metavolcanic rocks, quartzite, and metapsammite/schist (e.g., Hartville uplift, Snyder and others, 1989). This mixed sedimentary-volcanic association suggests that marine deposition (including very shallow water for the dolomite) in proximity to volcanic centers characterized at least part of the margin of the newly amalgamated Wyoming province during the Late Archean (Frost and Frost, 1993; Houston, 1993).



Figure 5. Aerial view of McIntosh Peak (right side of photograph in foreground), Moonstone Peak (left center of photograph), and surrounding Tertiary deposits, part of the western Granite Mountains (see Love, 1970). This area is part of the Sweetwater arch, a Laramide uplift that subsequently collapsed during late Cenozoic normal faulting (Sales, 1983; Hall and Chase, 1989) (see ***Figure 17*** *for an example of one of the late Cenozoic normal faults). McIntosh Peak and the surrounding mountains consist chiefly of Late Archean granitic rocks described by Peterman and Hildreth (1978) as typically massive and medium to coarse grained. Photograph by R.C. Thurston, Department of Botany, The University of Wyoming.*

The Wyoming province is one of seven Archean provinces that presently form Laurentia, the North American craton (Hoffman, 1988, 1989b). These Archean provinces could have been originally independent microcontinents, subsequently amalgamated in the Proterozoic. Alternately, they could have been part of an original Archean protocraton subsequently both rifted and reassembled during the Proterozoic (Hoffman, 1988). Consequently, relationships of the Wyoming province to its more extensive Canadian equivalents (e.g., Hearn and Superior provinces) are poorly known. Hoffman (1989b) noted that the Wyoming province differs from the Superior province by the presence, in the former, of abundant shelf-type metasedimentary rocks and isotopic data suggesting widespread continental crust older than 3.1 Ga (e.g., Wooden and others, 1988).

The Late Archean Wyoming province must have been larger than its present extent. Its present southern boundary, in part, reflects Early Proterozoic rifting and spreading that led to establishment of an Early Proterozoic passive margin (Karlstrom and others, 1983; Houston, 1993). Some mafic dike swarms within the Archean Wyoming province no doubt reflect this episode of crustal extension (Cox and others, 1995), although systematic geochronometric dating of mafic dike swarms in the Wyoming province is still in its infancy (Snyder and others, 1990).

An Early Proterozoic continental margin and its destruction

A fundamental Precambrian boundary is exposed in southeastern Wyoming, where the Early Proterozoic Colorado province of probable island-arc affinity was accreted to the Archean Wyoming craton and its Early Proterozoic passive margin cover (Hills and Houston, 1979; Karlstrom and Houston, 1984). This important tectonic boundary zone, originally referred to as the Mullen Creek-Nash Fork shear zone (Houston and McCallum, 1961), was renamed the Cheyenne belt by Houston and others (1979). This well-exposed

geosuture consists of a system of mylonite zones and varies in overall width from about 4.3 miles (7 km) to 0.4 mile (0.7 km) (Duebendorfer and Houston, 1986, 1987). In Wyoming, the event associated with the development of the Cheyenne belt is the oldest (ca. 1.76 Ga) in a complex Early Proterozoic crustal accretion history that added an enormous girth (>800 miles wide) of juvenile crustal material to the southwestern flank of nuclear North America during the interval 1790 to 1660 Ma (Condie, 1982; Nelson and DePaolo, 1985; Reed and others, 1987; Karlstrom and Bowring, 1988). The Cheyenne belt includes lenses of rocks of variable province affinity. The Bear Lake block of Duebendorfer and Houston (1987) includes orthogneiss of Archean affinity, whereas the Barber Lake block is composed of mixed provenance sediments yielding 2.0 to 2.4 Ga Nd model ages (Ball and Farmer, 1991). Nevertheless, south of the Barber Lake block no Archean rocks or components have been recognized in the extensive isotopic data reported by various workers (e.g., Nelson and DePaolo, 1985; Bennett and DePaolo, 1987; Ball and Farmer, 1991).

In the Medicine Bow Mountains, an exceptionally diverse and complete Early Proterozoic stratigraphic sequence is exposed north of the Cheyenne belt (Karlstrom and others, 1983; Houston and others, 1992). This sequence apparently records the rifting and eventual foundering of the Archean craton and development of an Early Proterozoic passive margin. Rifting is suggested by thick conglomeratic units of the Deep Lake Group inferred to be deposited in fault troughs floored by Archean basement (Houston, 1993). Intrusive mafic rocks in the Deep Lake Group have yielded a U-Pb zircon age of 2092 ±9 Ma (million years ago) (Premo and Van Schmus, 1989), providing the only geochronometric control for the inferred Early Proterozoic rifting event. An intriguing rock type in the Deep Lake Group is diamictite (Houston and others, 1981) interpreted as ancient glacial deposits (Blackwelder, 1926), based in part on the presence of dropstone clasts. These deposits have been correlated with the glaciogenic Gowganda Formation of the Huronian Supergroup of the Canadian Shield (Roscoe and Card, 1993). The Libby Creek Group which overlies the Deep Lake Group consists, in part, of an approximately 15,000-foot-thick (4.6-km-thick) clastic succession, which in turn is overlain by a 6500-foot-thick (2-km-thick) stromatolitic dolomite unit (Nash Fork Formation). Together, these units constitute a passive margin sequence, suggesting the classic transition from fluvial to shallow-marine clastic rocks to carbonate bank. The uppermost part of the Libby Creek Group, more problematic in origin, is characterized by locally pillowed basaltic rocks and graphitic slate. Such rocks clearly indicate a major change in the overall depositional framework and may signal the proximity of an allochthonous oceanic-arc terrane, as suggested by Houston (1993).

Rocks south of the Cheyenne belt are part of the Early Proterozoic Colorado province. They consist chiefly of various metamorphosed volcanogenic rocks, including both lavas and volcaniclastic rocks, complexly intruded by a variety of Proterozoic plutonic rocks ranging from mafic to felsic. In Wyoming, these rocks are exposed in the Sierra Madre, Medicine Bow Mountains, and Laramie Mountains (**Figure 2**). The best preserved volcanogenic rocks occur in the Sierra Madre, where they form part of the Green Mountain Formation of Divis (1976). Relict volcanic textures are preserved locally, allowing recognition of original lapilli tuff, volcanic breccia, and porphyritic phases. Geochemical studies of rocks from the Green Mountain Formation indicate an essentially bimodal basalt-dacite/rhyolite suite, with only sparse andesite, although an overall calc-alkaline trend is indicated (Schmidt, 1983; Condie and Shadel, 1984).

Zircons from a quartz-feldspar-mica gneiss (metadacite porphyry) of the Green Mountain Formation have yielded a U-Pb age of 1792 ± 15 Ma, representing the best constraint on the age of volcanism in the accreted arc terrane (Premo and Van Schmus, 1989). Some plutonic bodies south of the Cheyenne belt may be comagmatic with the metavolcanic rocks, whereas other Proterozoic plutons clearly are younger. The Sierra Madre Granite forms an important younger pluton, three phases of which have been dated as 1763 ± 6 Ma, 1749 ± 8 Ma, and 1744 ± 14 Ma by U-Pb zircon techniques (Premo and Van Schmus, 1989). This pluton intruded a mylonite zone related to the plastic deformational history of the Cheyenne belt but is itself cut by a brittle cataclastic fault zone. Houston (1993) argued that this granitoid constrains the final stages of the suturing process between the Colorado and Wyoming provinces as manifested in the Sierra Madre. Therefore, the deformational age of the Cheyenne belt is broadly defined as post-1790 Ma but pre-1750 Ma.

The Cheyenne belt has not been conclusively identified east of the Medicine Bow Mountains principally because the Middle Proterozoic Laramie Anorthosite Complex and portions of the associated Sherman batholith intrude the inferred extension of the suture zone in the Laramie Mountains. However, east of the Laramie Mountains, a mylonitic zone in the Richeau Hills may be a manifestation of the Cheyenne belt

(Duggan and others, 1995; Harper and Chamberlain, 1996 and unpublished data). Nevertheless, a much more extensive and better exposed Early Proterozoic mylonitic shear zone occurs in Archean rocks of the central Laramie Mountains (i.e., Laramie Peak shear zone of Chamberlain and others [1993]). Sphene from an Early Proterozoic mafic dike deformed in this mylonitic shear zone has yielded an age of 1763 ± 7 Ma and is thus within the age bracket for Cheyenne-belt deformation defined in the Medicine Bow Mountains and Sierra Madre (Resor and others, 1996). The Laramie Peak shear zone is interpreted as the frontal fault zone of a block uplift of tectonically reactivated Archean rocks regenerated during the development of the Cheyenne belt.

The northern boundary of the Wyoming province is still poorly understood but apparently is manifested by the reactivated Great Falls tectonic zone of east-central Idaho and west-central Montana (O'Neill and Lopez, 1985; O'Neill and Harlan, 1992). The buried Precambrian rocks north of this subsurface crustal discontinuity have been referred to as the Medicine Hat block (Hoffman, 1989b, figure 12). Proterozoic reactivation of Archean basement of the northwestern Wyoming province has been demonstrated by Erslev and Sutter (1990) in the southern Madison Range (Montana) as manifested by the 3-km-thick Madison mylonite zone. These authors hypothesized that the Madison mylonite zone was an Early Proterozoic (1.8 to 1.9 Ga or possibly later) foreland thrust zone that developed on the margin of a major contractile orogen. The Archean Wyoming province thus has experienced significant Early Proterozoic reactivation along its northwestern margin as well as southeastern margin.

1.4 to 1.5 Ga magmatism — anorogenic or syntectonic?

A transcontinental magmatic event that occurred about 1.4 to 1.5 Ga has been recognized from Labrador to southern California (Silver and others, 1977; Anderson, 1983). In southeastern Wyoming, this magmatism is well represented by the Laramie Anorthosite Complex (LAC) and Sherman batholith (Frost and others, 1993). The LAC was emplaced across the inferred projection of the Cheyenne belt into the south-central Laramie Mountains (**Figures 2** and **4**). The youngest pluton of the complex (Red Mountain pluton) has yielded a U-Pb zircon age of 1439 +7/-6 Ma (Frost and others, 1990). Geochronometric studies on granitic rocks from the Sherman batholith indicate a crystallization age of about 1.43 Ga (Zielinski and others, 1981; Aleinikoff, 1983), thereby suggesting the broadly coeval nature of these plutonic suites. Field relationships, however, indicate that anorthositic rocks of the LAC predate the granitic rocks of the Sherman batholith, whereas the late monzosyenitic rocks of the LAC locally grade into granitic rocks of the Sherman batholith (Frost and others, 1993).

The LAC is a multi-stage, lithologically heterogeneous plutonic complex. Anorthositic rocks ranging from pure anorthosite to plagioclase-rich gabbro, norite, and troctolite are the main components of the complex [see Frost and others (1993) for exact petrographic nomenclature]. Four domical structures, as defined by distribution and attitude of foliation, are present in the anorthositic rocks. Foliations in the anorthositic rocks range from primary igneous layering and plagioclase lamination to subsolidus flow layering, probably related to the forceful upward emplacement of a hot, but nearly solidified, aggregate of crystals [see Frost and others (1993) for specific details]. Various monzosyenitic to granitic rocks form discrete younger plutons along the margins of the composite anorthositic body.

The Sherman batholith is composed chiefly of potassic, iron-rich (relative to MgO) granitic rocks. The dominant granitic rock is massive megacrystic, biotite-hornblende granite. Other granitic rocks comprising the batholith (Frost and others, 1993) have not been delineated systematically on a geologic map. A suite of fine-grained granitoids forms part of the Sherman batholith, and Frost and others (1993) suggested that these rocks may be chilled equivalents of the granite. Near Virginia Dale, Colorado, Eggler (1968) documented a ring-dike complex as part of the Sherman batholith, suggesting an overall epizonal emplacement history for this mid-Proterozoic pluton.

The widespread 1.4 to 1.5 Ga magmatism of North America commonly is referred to as anorogenic (e.g., Anderson, 1983), implying that pluton emplacement was not associated with a specific orogenic event such as subduction or collision. This petrotectonic classification is based principally on the observation that Proterozoic deformation and metamorphism younger than 1.65 Ga has not been found associated with the transcontinental belt (Anderson and Bender, 1989; but see below). Furthermore, various levels of emplacement are exposed along the belt, so that the anorogenic characterization is not simply an artifact of epizonal plutonic emplacement (Anderson and Bender, 1989). Windley (1993) argued that the 1.4 to 1.5 Ga magmatism was a manifestation of the breakup of a supercontinent, and thus developed during an exten-

sional regime. In contrast, Nyman and others (1994) recommended abandoning the term "anorogenic" and argued that deformation in the 1.4 to 1.5 Ga plutons, albeit highly partitioned, was related to compressional or transpressional plate-margin tectonism along the southern margin of Laurentia. This model requires that portions of Laurentia, thermally softened by the Middle Proterozoic magmatic intrusions, were deformed >1000 km inboard from the southern plate margin. Finally, Hoffman (1989a) proposed the hypothesis of a "mantle superswell"–convective upwelling, thousands of kilometers in diameter, beneath a stationary supercontinent that aggregated previously in the late Early Proterozoic. According to this concept, mantle magmas invaded and ponded in the lower crust, causing partial melting and uplift. Rifting, if it occurred, would have been a consequence, not the cause, of the mantle upwelling (Hoffman, 1989a). Regardless of the tectonic setting of 1.4 to 1.5 Ga magmatism, a remarkable continent-wide event occurred during the Middle Proterozoic. Juvenile Proterozoic crust experienced deep-seated crustal melting, eventually leading to widespread distribution of potassic (rapakivi-type) granitic complexes with the local association, especially in Canada but also in Wyoming, of anorthosite and monzosyenitic rocks.

Missing Proterozoic record

Between the Middle Proterozoic magmatic event and the beginning of Paleozoic sedimentation on the Wyoming craton, there is no record of nearly 1 billion years of Earth history within the confines of the state. A few mafic dikes have yielded radiometric dates within this time span [see Snyder and others (1990), figure 9.15] and may constitute an exception. Nevertheless, the long interval presumably represents emergence and erosion of this part of Laurentia prior to the Late Proterozoic breakup of the supercontinent (Hoffman, 1991).

Some evidence for the general Proterozoic tectonic framework of the region, however, is preserved in areas adjacent to Wyoming. During the Middle Proterozoic, the Belt-Purcell Supergroup of the Northern Rocky Mountains accumulated in the "Belt basin" as a thick stratigraphic unit [perhaps up to 52,500 feet (16 km), Winston (1989)] consisting chiefly of siltite, argillite, quartzite, and carbonate rocks (Harrison, 1972). The tectonic setting of the Belt basin is controversial, but one model suggests an intracratonic rift (Winston, 1986). Northwest-striking normal faults related to development of the Belt basin have been recognized in southwestern Montana and similar faults probably extended into northwestern Wyoming (Schmidt and Garihan, 1986b, figure 7). During deposition of the Belt-Purcell Supergroup, Wyoming apparently was a faulted upland. Nd-Sm isotopic data (Frost and Winston, 1987), however, suggest that rocks of the Archean Wyoming province were not the prime source of the terrigenous sediment that accumulated in the Belt basin.

Immediately south of the southwestern border of Wyoming, in northeastern Utah and adjacent Colorado, the approximately 24,000-foot-thick (7 km) Uinta Mountain Group of terrigenous sandstone (and quartzite), shale, and conglomerate was deposited in an east-west striking, fault-bounded basin (inferred aulacogen; e.g., Burke and Dewey, 1973) during the Middle to Late Proterozoic (Hansen, 1965). The maximum age range of the Uinta Mountain Group is 770 to 1110 Ma, which is generally younger than the inferred age range of the Belt and Purcell Supergroups of the Northern Rocky Mountains [Bressler, 1981; see Winston (1989) for a summary of the age of the Belt]. Localization of the inferred Uinta aulacogen along the trace of the extension of the Cheyenne belt (Sears and others, 1982) suggests tectonic heredity and the importance of this suture zone to the post-Early Proterozoic structural development of the Rocky Mountain region (Bryant, 1985). Furthermore, similar east-west structural trends in the Wyoming foreland may reflect a Precambrian structural grain that was established in Archean basement rocks during the Proterozoic and reactivated during later tectonism (Bryant, 1985).

The rifting that eventually led to the breakup of the supercontinent Laurentia was part of a protracted rift history that began about 730 to 770 Ma (Devlin and Bond, 1988). Deposition of the Upper Proterozoic Windermere Supergroup (or equivalents) represents early manifestations of this rift history, and these sedimentary rocks occur as thick stratigraphic sequences west of Wyoming. Although structures related to this long history of crustal extension may exist in Precambrian basement rocks of Wyoming, no such feature has been definitely recognized. However, Mitra and Frost (1981) argued that retrograde deformation zones containing chlorite and actinolite, formed between 900 and 600 Ma, are common throughout Precambrian rocks of the Wind River Range.

In summary, between the Middle Proterozoic and the beginning of the Phanerozoic, Wyoming apparently was a positive area that may have periodically experienced extensional faulting. Large areas adjacent to Wyoming were significantly extended, eventually leading to the development of deep basins and the

accumulation of thick sedimentary successions. In some cases, the area of paleo-Wyoming served as a significant source area for terrigenous sediments that accumulated in these basins.

Early and middle Paleozoic shelf-facies sedimentation

At the dawn of the Paleozoic Era, Wyoming was situated on the west-central flank of rifted Laurentia (Hoffman, 1988). The complex Precambrian basement of Wyoming served as a low-relief platform for accumulation of a thin cratonic sedimentary cover during the early to middle Paleozoic (Boyd, 1993). The Transcontinental arch was a major paleotectonic feature that influenced the depositional history on the Rocky Mountain shelf especially during the Paleozoic Era. This feature was a broad, somewhat sprawling, southwest-northeast-trending basement high that extended from present-day northeastern Arizona to southern Minnesota and perhaps across the Canadian shield. Wyoming's Phanerozoic depositional record therefore accumulated on the northwestern flank of this large-scale basement arch. Influence of the Transcontinental arch is indicated by stratigraphic thickness and facies variations associated with the progressive west-to-east marine transgression that deposited the Sauk sequence during the Cambrian and Early Ordovician (Sloss, 1963; Lochman-Balk, 1972). However, its role as a long-lived (reactivated) paleotectonic feature is especially obvious in the absence of Silurian beds over a broad area of the west-central United States (Gibbs, 1972; Peterson and Smith, 1986). Paleogeographic reconstructions suggest that marine Middle Silurian strata covered the craton (Dott and Batten, 1981). Furthermore, definitive evidence of the presence of Middle Silurian rocks in southeastern Wyoming is provided by fossiliferous blocks of sedimentary rocks in diatremes near the Wyoming-Colorado border (Chronic and others, 1969). Outside of the diatreme occurrences, the distribution of Middle Silurian rocks suggests broad-scale Late Silurian to Early Devonian uplift of the Transcontinental arch, associated with widespread erosion (the development of the pre-Kaskaskia unconformity).

The oldest known Phanerozoic stratigraphic unit in Wyoming is a quartz-rich sandstone (Flathead Sandstone) that formed an eastward-transgressive sheet across Wyoming during the Middle to Late Cambrian. The basal contact of this unit is the fundamental "great unconformity" seen between Precambrian and Phanerozoic rocks (base of the Sauk sequence of Sloss, 1963). Farther to the west, in the site of the developing Cordilleran miogeocline, an enormous thickness of Upper Precambrian and Lower Cambrian strata had accumulated along the rifted continental margin (Stewart, 1972). Wyoming, however, lay east of a regional hinge line (the so-called Wasatch line of Utah) and correlative deposits are absent on the Wyoming shelf. An enormous discrepancy in thicknesses of broadly coeval units originally deposited east and west of the Wasatch line continued as an essential characteristic of the Paleozoic depositional history.

On the Wyoming craton, the Flathead Sandstone is chiefly overlain by various thin, shallow-water marine deposits; disconformities are common within this stratigraphic succession. However, during the Early Devonian, much of Wyoming was emergent, and only lenticular estuarine deposits are preserved locally (e.g., Beartooth Butte Formation in the Beartooth Mountains of northwestern Wyoming and southwestern Montana). The Mississippian was a time of extensive carbonate deposition throughout Wyoming as well as over a large part of interior North America (Dott and Batten, 1981). In Wyoming these carbonate rocks are commonly dolomitic. Various stratigraphic names have been applied around the state (Boyd, 1993; Love and others, 1993); throughout central Wyoming, these rocks are called the Madison Limestone (Lageson and others, 1979). The Mississippian carbonate strata vary from a zero edge in southeastern Wyoming to more than 1000 feet (305 m) thick along the western and northern borders of the state (Peterson and Smith, 1986).

Widespread emergence of much of the continental interior at the close of deposition of the Madison Limestone caused extensive erosion and dissolution of its upper surface. As a result, solution-breccia beds and paleokarst features are common in upper parts of the Madison Limestone [see Huntoon (1993) for a discussion of the hydrologic significance]. This unconformity coincides with the top of the Kaskaskia sequence of Sloss (1963). Overlying units such as the Amsden and Tensleep formations are primarily clastic rocks, indicating fundamental changes in depositional and tectonic conditions of the western North American craton (Peterson, 1988).

Late Paleozoic and early Mesozoic depositional basins

During Pennsylvanian through Permian time, the southern margin of North America experienced a major collisional event that is in part manifested by the Ouachita-Marathon orogenic belt. An aspect of this

orogeny was reactivation of pre-existing zones of weakness and uplift of large crustal blocks in the foreland of the orogenic belt (Kluth and Coney, 1981). This intraplate deformation can be traced across a broad belt that extends west-northwest from the structural front of the Ouachita-Marathon orogenic belt to the continental margin of late Paleozoic North America as seen in Nevada and Utah (Kluth, 1986). Tectonic development of the Ancestral Rocky Mountains is a manifestation of this foreland deformation. Associated uplifts and their basins formed principally in Colorado, but important extensions affected parts of southeastern Wyoming (Maughan, 1993).

In southeastern Wyoming, the Ancestral Rocky Mountains were manifested principally by the Pathfinder uplift, originally delineated by Mallory (1963). However, the extreme northern end of the Ancestral Front Range uplift also extended into Wyoming near the present site of the southwestern Medicine Bow Mountains and Sierra Madre (Mallory, 1963; Maughan, 1993). Delineation of uplifts associated with the Ancestral Rocky Mountains in Wyoming is based on: (1) stratigraphic relationships presently preserved on the flanks of younger, superposed Laramide uplifts such as the Laramie Mountains and Sweetwater arch, and (2) subsurface data from adjacent Laramide basins [especially see Mallory (1967)]. An important unit in the delineation of these ancient uplifts is the widespread Amsden Formation, which ranges in age from Late Mississippian to Middle Pennsylvanian [Sando and others, 1975; but see Boyd (1993) for a discussion of the "Amsden problem"]. Mallory (1963) used the absence of Amsden Formation in southeastern Wyoming to define the extent of the Pathfinder uplift. In this area, the Amsden Formation probably was chiefly Morrowan and Atokan (Mallory, 1967) and, consequently, its absence indicates syn- or post-Atokan erosion.

Mallory (1963, 1967) also demonstrated that upper parts of the Amsden Formation (Atokan age) systematically thicken away from the inferred uplift, suggesting that the uplift was of Atokan age. Southern parts of the Pathfinder uplift apparently were eroded to Precambrian basement during the Middle Pennsylvanian. For example, along the west flank of the southern Laramie Mountains, south and east of Laramie, Precambrian basement rocks are immediately overlain by red sandstone, conglomerate, and arkose of the Fountain Formation (Knight, 1929), which ranges in age from Atokan to Virgilian. Lower Paleozoic rocks are absent in the southern Laramie Mountains, although regional paleogeographic reconstructions and fossiliferous blocks of Ordovician and Silurian rocks in diatremes near the Wyoming-Colorado border suggest their former presence in the area (Chronic and others, 1969).

During the Permian, Wyoming was part of the western margin of the supercontinent Pangaea, and was situated near the equator (Smith and others, 1981). A regional disconformity in central Wyoming separates Pennsylvanian and Permian strata. In contrast, along the western and eastern margins of the state, a more continuous Upper Pennsylvanian to Lower Permian record is preserved (e.g., the Wells Formation in the west and Casper and Minnelusa formations in the east). Permian stratigraphy in Wyoming is characterized by the intertonguing of numerous lithologic units that has fostered a bewilderingly complex stratigraphic nomenclature (Peterson, 1984; Boyd, 1993).

In the eastern part of the state, a redbed and evaporite facies, the Goose Egg Formation, is common, whereas shelf carbonates are important components of the Phosphoria (Park City) Formation in the central to western parts of the state. Phosphatic shale, phosphorite, and bedded chert are distinctive components of the Phosphoria Formation of western Wyoming. The organic-rich members of the Phosphoria Formation are especially significant as source rocks for hydrocarbons and as a phosphate reserve. The top of the Permian System exhibits an erosional surface across the Cordilleran miogeocline; Pipiringos and O'Sullivan (1978) suggested development of a general unconformity (their R-1 unconformity) at the Permian-Triassic boundary on the stable shelf of the Western Interior.

Red colors are commonly associated with Triassic strata in Wyoming. As Picard (1993) vividly described, the redbeds of the Chugwater Group (of Pipiringos, 1968) are visually the most spectacular stratigraphic units in Wyoming. During Chugwater deposition, much of the state was a virtually featureless, muddy coastal plain. However, coeval marine siltstones and limestones accumulated in western Wyoming in the form of the Dinwoody and Thaynes formations. The Alcova Limestone is a distinctive marker unit in the Chugwater Group. This laminated, purplish gray, resistant limestone has been interpreted as a marine tongue extending eastward from upper parts of the Thaynes Formation into the redbed sequence (Picard and others, 1969).

After the Early Triassic, marine waters temporarily withdrew from the U.S. Cordilleran miogeocline (Carr and Paull, 1983). Regional emergence and erosion occurred, manifested by the widespread R-3 unconformity of Pipiringos and O'Sullivan (1978). For

example, on the Colorado Plateau, the Tr-3 unconformity separates the Upper Triassic Chinle Formation from the underlying Lower Triassic Moenkopi Formation. However, in Wyoming the unconformity is much more subtle; according to Pipiringos and O'Sullivan (1978), Tr-3 occurs within the Crow Mountain Sandstone and Jelm Formation of the Chugwater Group. The Popo Agie Formation of Wyoming's Chugwater Group (Picard, 1993) is temporally equivalent to continental beds of Arizona's Chinle Formation.

The eolian Nugget Sandstone, which is widespread in western and south-central Wyoming, is a distinctive and paleoenvironmentally significant Lower Jurassic unit. The regional J-0 unconformity of Pipiringos and O'Sullivan (1978) is at the base of the Nugget Sandstone. The Nugget Sandstone was part of an enormous Early Jurassic coastal to inland dune field that stretched from central Wyoming to southern Arizona (Kocurek and Dott, 1983). Other formations that were part of this great eolian sand sea include the scenic Navajo Sandstone of southern Utah and northern Arizona and the Aztec Sandstone of southern Nevada and adjacent California. In southern Arizona and southeastern California, equivalent eolian deposits intertongue with volcanic rocks of an Early Jurassic magmatic arc (Busby-Spera, 1988). The Nugget Sandstone is a profilic reservoir for hydrocarbons in the Wyoming-Utah part of the Cordilleran fold-and-thrust belt (Picard, 1993).

Marine waters returned to Wyoming and the craton region in the Middle Jurassic with establishment of a shallow sea (Sundance Sea) that apparently extended from Canada to northern Arizona (Kocurek and Dott, 1983). This transgression was an early stage of a gradual worldwide rise in sea level that eventually would reach its zenith in the Late Cretaceous (Vail and others, 1977). In Wyoming, various rock units manifest this major transgression, including the Gypsum Spring Formation, Twin Creek Limestone, and Sundance Formation. These units represent the initial deposits of the classic Zuni sequence of Sloss (1963). The J-2 unconformity of Pipiringos and O'Sullivan (1978), which extends throughout the Western Interior, is at the base of this transgressive sequence.

During the Late Jurassic, a major interruption of the Jurassic transgressive sequence occurred throughout the Rocky Mountain region. Beginning approximately in the mid-Oxfordian, the Rocky Mountain shelf was transformed into a nonmarine environment, with deposition of the widespread, chiefly fluvial, Morrison Formation. Although global sea level was continuing to rise at this time (Vail and others, 1977), a significant increase in siliciclastic debris in the Rocky Mountain region, probably related to increased orogenic activity to the west (e.g., Allmendinger and others, 1984; Thorman and others, 1990), filled the Jurassic seaway (Brenner, 1983). The Rocky Mountain shelf gradually became covered with varicolored continental deposits characterized by complex intertonguing of mud, sand, gravel, and lenses of lacustrine limestone. Volcanic ash, derived from distant volcanic centers to the west, is a conspicuous detrital component in these nonmarine rocks. Spectacular dinosaur fossils have been collected from the Morrison Formation at several localities in the Rocky Mountains; but at Como Bluff, Albany County, southeastern Wyoming, this paleontologically significant unit has also yielded the greatest diversity of Mesozoic mammals in the world (Clemens and others, 1979). The Morrison Formation throughout the Rocky Mountain region is bounded at its base by the J-5 unconformity of Pipiringos and O'Sullivan (1978), and its top is usually beveled by another unconformity.

Western Interior Cretaceous basin

At its maximum development, the Western Interior Cretaceous basin extended from the Arctic Ocean to the Gulf of Mexico (Kauffman, 1977). This epicontinental sea was separated from the Pacific Ocean by Cordilleran highlands, which also served as the main provenance of sediment shed eastward into the basin. In the Rocky Mountains, a regional unconformity of uncertain duration usually separates middle Cretaceous rocks, including widespread conglomerates (Heller and Paola, 1989), from the Upper Jurassic fluvial deposits of the Morrison Formation. This hiatus is synchronous with the Early Cretaceous magmatic lull that is well documented throughout the western North American Cordillera (Armstrong and Ward, 1993). By contrast, synchronous with the culmination of seaway advancement was widespread mid- to Late Cretaceous magmatism (Armstrong and Ward, 1993) as well as the complex shortening history of the foreland fold-and-thrust belt. Thicknesses of late Lower to Upper Cretaceous strata increase dramatically across Wyoming, from about 6000 to 7000 feet (1.8 to 2.1 km) in northeastern Wyoming to more than 15,000 feet (4.6 km) in the southwest (Peterson and Smith, 1986) (**Figure 6**). Evolution of the Western Interior Cretaceous basin is chronologically linked with orogenesis along its western margin (Jordan, 1981; Steidtmann, 1993). Coincident with tectonic loading of the foreland fold-and-thrust belt during the mid-

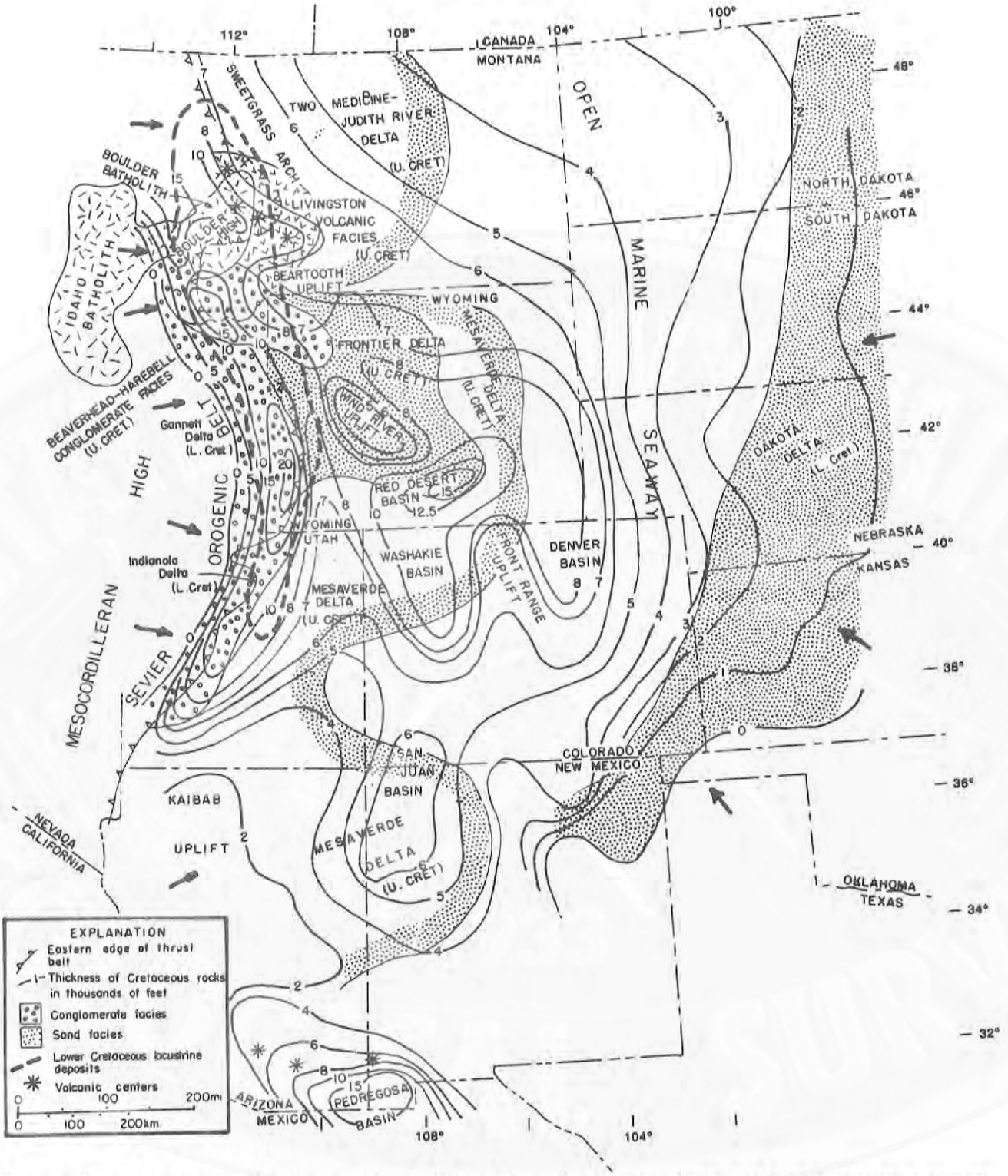


Figure 6. Cretaceous System, showing approximate thickness, general sedimentary facies, and main paleotectonic elements. Arrows indicate probable transport direction of terrigenous clastic sediments (from Peterson and Smith, 1986, figure 12; used with permission).

Cretaceous was a worldwide rise of sea level (Vail and others, 1977; Hallam, 1984), which amplified the development of this broad seaway.

Rocks of the Cretaceous Western Interior basin have been major sources of oil, gas, and coal in the Rocky Mountain region. These economic aspects, coupled with excellent exposures along the margins of many Laramide basins, have encouraged detailed analysis of these rocks. Correlation between isolated exposures is made possible by the presence of widespread and distinctive fossil assemblages (especially ammonites) and abundant subsurface data. These

Cretaceous rocks are chiefly shale, siltstone, and sandstone. Conglomerate is also important along western margins of the basin, and limestone is generally restricted to the east (see McGookey and others, 1972, for various detailed stratigraphic sections). An important component of the fine-grained sedimentary rocks of the Western Interior Cretaceous basin is volcanic ash (now mostly bentonite). The abundance of bentonitic beds in these deposits is a by-product of extensive, coeval arc magmatism along the western margin of North America throughout the middle to Late Cretaceous (Kauffman, 1977; Hamilton, 1988a).

Complex facies relationships exist between these various Cretaceous rocks, including intertonguing of nonmarine and marine deposits as the shoreline transgressed and regressed (Steidtmann, 1993). The geological significance of the migration of the shoreline is controversial. Some authors interpret these variations as chiefly related to global sea-level changes (Hancock and Kauffman, 1979), whereas others argue that local or subregional tectonic events were the principal causes (Lillegraven and Ostresh, 1990). Recognizing and dating intrabasinal unconformities are aspects in evaluating the relative importance of global sea-level changes versus localized tectonics in the evolution of depositional systems of the Western Interior basin. At least nine regional unconformities have been identified. These unconformities are completely within nonmarine strata, involve both marine and nonmarine rocks, or are totally within marine strata (Weimer, 1984). At least three of these unconformities (at 97, 90, and 80 Ma) appear to be primarily related to drops in sea level; however, either sea-level changes or tectonics could have been responsible for the other unconformities (Weimer, 1984). Many minor, short-term variations in strandline position, especially along the western margin of the basin, undoubtedly were influenced by local tectonic events.

Fold-and-thrust belt of the Sevier orogeny

Paleogeographic setting

Late Cretaceous paleogeography of western North America was dominated by three sinuous mountain belts that extended virtually the length of the Cordillera: (1) a thin-skinned fold-and-thrust belt on the east; (2) a central hinterland; and (3) an Andean-type, volcanic-plutonic magmatic arc on the west [see Miller and others (1992) for a regional summary]. The fold-and-thrust belt on the east was localized at the miogeocline-to-craton hinge line. Initial eastward thrusting of thicker miogeoclinal strata over thinner cratonic strata was followed by progressively more eastward-propagating thrusts (Armstrong and Oriel, 1965; Royse and others, 1975). The roots of this thrust belt pass westward into the hinterland, where extensive tracts of metamorphic rocks and even Precambrian basement rocks are involved in the Mesozoic deformation (Lush and others, 1988) (**Figure 7**). The hinterland is a zone of Mesozoic crustal thickening, manifested by its Jurassic and Cretaceous metamorphic, magmatic, and deformational features (Coney and Harms, 1984; Snoke and Miller, 1988). Some geologists have suggested that the Late Cretaceous hinterland was analogous to the Cenozoic Andean Altiplano (e.g., Allmendinger, 1986), which is a high, arid plateau situated between the magmatic arc of the western Cordillera and the fold-and-thrust belt of the eastern Cordillera. The Mesozoic magmatic arc is characterized by an extensive batholithic belt extending

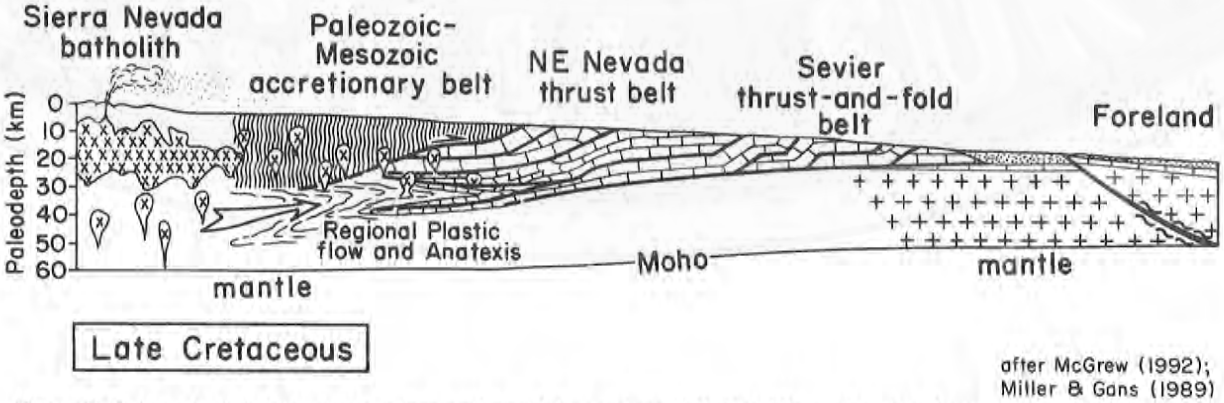


Figure 7. Schematic late Cretaceous cross-section of the western U.S. Cordillera, showing the relationship between the Sevier fold-and-thrust belt and fractured Rocky Mountain foreland on the east, metamorphic and igneous rocks of the hinterland, and the Sierra Nevada batholith on the west. Adapted from McGrew (1992) and inspired from Miller and Gans (1989, figure 4).

from Baja California to the Coast Plutonic Complex of western British Columbia and southeastern Alaska. Magmatism in this batholithic belt spanned the interval from about 225 Ma to 55 Ma, but reached a culmination during the mid- to Late Cretaceous (125 to 85 Ma, Armstrong and Ward, 1993).

Timing and structural style

The imbricate thrust belt exposed in western Wyoming, southeastern Idaho, and northern Utah is a major salient in the Cordilleran fold-and-thrust belt (**Figure 1**). This salient of the fold-and-thrust belt has been the site of classic geologic studies (e.g., Veatch, 1907; Schultz, 1914; Rubey and Hubbert, 1959; Armstrong and Oriel, 1965; Royse and others, 1975) and extensive hydrocarbon exploration. The thrust belt is generally not deeply eroded and thus provides excellent exposures of contractile structures at a shallow level in the crust. This segment of the Cordilleran fold-and-thrust belt is one of the best understood thin-skinned, foreland fold-and-thrust belts in the world (Royse, 1993; Lageson and Schmitt, 1994).

The Wyoming-Idaho-Utah fold-and-thrust belt, part of the Sevier orogenic belt (Armstrong, 1968), is a classic example of an intraplate, retro-arc fold-thrust belt. Contractile deformation manifested in this belt apparently began during the Early Cretaceous (Heller and others, 1986), although earlier studies (e.g., Armstrong and Oriel, 1965; Wiltschko and Dorr, 1983) suggested a possible Late Jurassic initiation. The Wyoming-Idaho-Utah fold-thrust belt also includes an exceptional record of synorogenic sedimentary rocks that date progressive thrust fault development and displacement during sedimentation (e.g., Armstrong and Oriel, 1965; Royse and others, 1975; Wiltschko and Dorr, 1983; Steidtmann and Schmitt, 1988; Lageson and Schmitt, 1994). Progressive incorporation of early synorogenic sedimentary strata into the advancing thrust wedge, coupled with the overlap of fault traces by younger sedimentary deposits, bracket the timing of motion on many individual thrust sheets (e.g., Armstrong and Oriel, 1965).

The geometry and style of major structures in the Wyoming-Idaho-Utah fold-and-thrust belt are well documented (Royse and others, 1975; Lamerson, 1982; Royse, 1993). Structures characteristic of the Wyoming portion of the fold-and-thrust belt are basically analogous to the "Foothills family" of structures, as recognized by Dahlstrom (1970) for the Canadian Rocky Mountains (see Lageson and Schmitt, 1994, for a summary, their table 1). The thrusts are thin-skinned, affecting only sedimentary rocks and merging into a décollement near the top of crystalline basement. They dip to the west, are listric in shape, and cut up-section in the direction of transport in a stair-step trajectory (**Figure 8**). Glide planes are long and flat and localized within weak layers such as Cambrian shale, Jurassic salt, and Cretaceous shale. Ramps are short and cut across more competent rocks such as Mississippian carbonate or Jurassic sandstone (Platt and Royse, 1989). Folds are generated chiefly by bending above ramps and commonly are kink-like in style (Boyer, 1986). Mylonitic rocks generally are absent along the thrust surfaces. Bedding-parallel shortening features (e.g., solution cleavage) are characteristic of parts of the thicker western thrust sheets (Mitra and others, 1988).

Models for the Sevier orogeny

Tectonic models for the Sevier orogeny (the retro-arc or foreland fold-and-thrust belt) are numerous and controversial. Early models for parts of the Cordilleran fold-and-thrust belt that involve gravitational gliding (e.g., Crosby, 1968; Mudge, 1970; Scholten, 1973) have been demonstrated to be untenable. Principal problems involve the requirement for significant slope; the downward rooting of the thrust faults, as indicated by detailed, regional cross sections; and the absence of a synchronously denuded area that could have compensated for imbrication in the thrust wedge (Price, 1971). Lateral gravitational spreading related to buoyant upwelling of metamorphic infrastructure, as envisioned by Price and Mountjoy (1970), has been abandoned by some workers because of apparent problems related to detailed timing of metamorphism in the infrastructure and shortening in the foreland. However, there is a growing body of evidence that suggests a broad synchroneity between magmatism, metamorphism, and contraction in the hinterland (e.g., northeastern Nevada) and shortening associated with at least the Late Cretaceous phase of the Sevier fold-and-thrust belt (Miller and Gans, 1989; Camilleri and Chamberlain, 1997) (**Figure 7**).

Hamilton (1978) and Smith (1981) argued for a close connection between arc magmatism and thrusting. They envisioned that thickening and uplift of the crust during arc magmatism developed a topographic high (i.e., arc inflation of the crust), which served to drive foreland thrusting. The detailed timing between arc magmatism and thrusting in the Sevier fold-and-thrust belt remains controversial with some workers stating that thrusting along the eastern margin of the belt (e.g., central Utah) is significantly younger (10 to 20 m.y.) than the youngest plutonism at similar lati-



Figure 8. A part of the eastern front of the western Wyoming fold-and-thrust belt [i.e., Darby thrust system of Royse (1993)] is exposed at Battle Mountain (left center of photograph), Hoback River Canyon. The view is looking northward from the west bank of the Hoback River (in foreground). At this locality, Jurassic-Triassic strata (most prominently the Jurassic(?)-Triassic(?) Nugget Sandstone in this photograph) form the hanging wall of the Cliff Creek thrust of Dorr and others (1977), whereas vertebrate fossil-bearing sandstone and claystone of the Paleocene Hoback Formation form the footwall (grass- and tree-covered hills, right center of photograph). However, Hunter (1988) demonstrated with subsurface data and cross section construction that a deeper, blind thrust fault (his "Granite Creek thrust" and related faults) defines the actual true front of the fold-and-thrust belt in this area and is part of a triangle zone that also involves the east-dipping Game Hill fault of Dorr and others (1977). Photograph by A.W. Snoke.

tudes (Miller and others, 1992, p. 223). Warren Hamilton (personal communication, 1994) argued that crustal spreading related to arc inflation would continue after the cessation of arc magmatism, and thus only a broad correlation from thrusting to magmatism is required in his model. The synthesis study of Cordilleran magmatism by Armstrong and Ward (1993) has demonstrated two major culminations of magmatism and convergence–one in the Middle to Late Jurassic and another in the mid- to Late Cretaceous. Both culminations correspond to one of the two major depositional cycles of the Mesozoic Western Interior basin and to eustatic sea-level maxima, but only the Cretaceous culmination is synchronous with widespread foreland thrusting. Finally, intracontinental subduction (so-called Ampferer or A-type subduction, Bally, 1975; Scholten, 1982) involves westward subduction or shortening of cratonic crust with concomitant imbrication of the sedimentary cover along a décollement system that deepens westward. Various forms of this model have been proposed (e.g., Misch, 1960; Sales, 1968; Burchfiel and Davis, 1968, 1975; Campbell, 1973). Armstrong and Ward (1993) argued that high heat flow behind the mid- to Late Cretaceous Andean magmatic arc, which stretched from Mexico to Alaska, thinned the lithosphere under miogeoclinal areas and facilitated imbrication of the crust during overall trench-continent convergence. An outgrowth of the intracontinental subduction model is the concept of the "orogenic float," which assumes a fundamental decoupling between the complex structures of the crust and simpler lithospheric roots (Oldow and others, 1989a, 1990). This hypothesis argues for complete allochthoneity of crust across the whole of the Cordilleran orogen.

Laramide orogeny

During the Late Cretaceous and early Tertiary, the Rocky Mountain foreland was fractured by deep-rooted, reverse and thrust faults that uplifted broad blocks of Precambrian basement rocks, separated by deep basins (**Figure 9**). This thick-skinned style of deformation is characteristic of the classic "Laramide orogeny" of specific tracts within the Rocky Mountain region. Considerable debate continues on precise timing of this fracturing and crustal shortening of the foreland, but recent studies suggest the interval 75 to 45 Ma (Cross, 1986; Dickinson and others, 1988). In Wyoming, however, stratigraphic and structural data indicate that the Laramide orogeny ended by 51 Ma (Lillegraven, 1993). Some authors argue that Laramide-related orogenic events in Wyoming began earlier (e.g., Steidtmann and Middleton, 1991) and continued later (e.g., Steidtmann and others, 1989) than this time span.

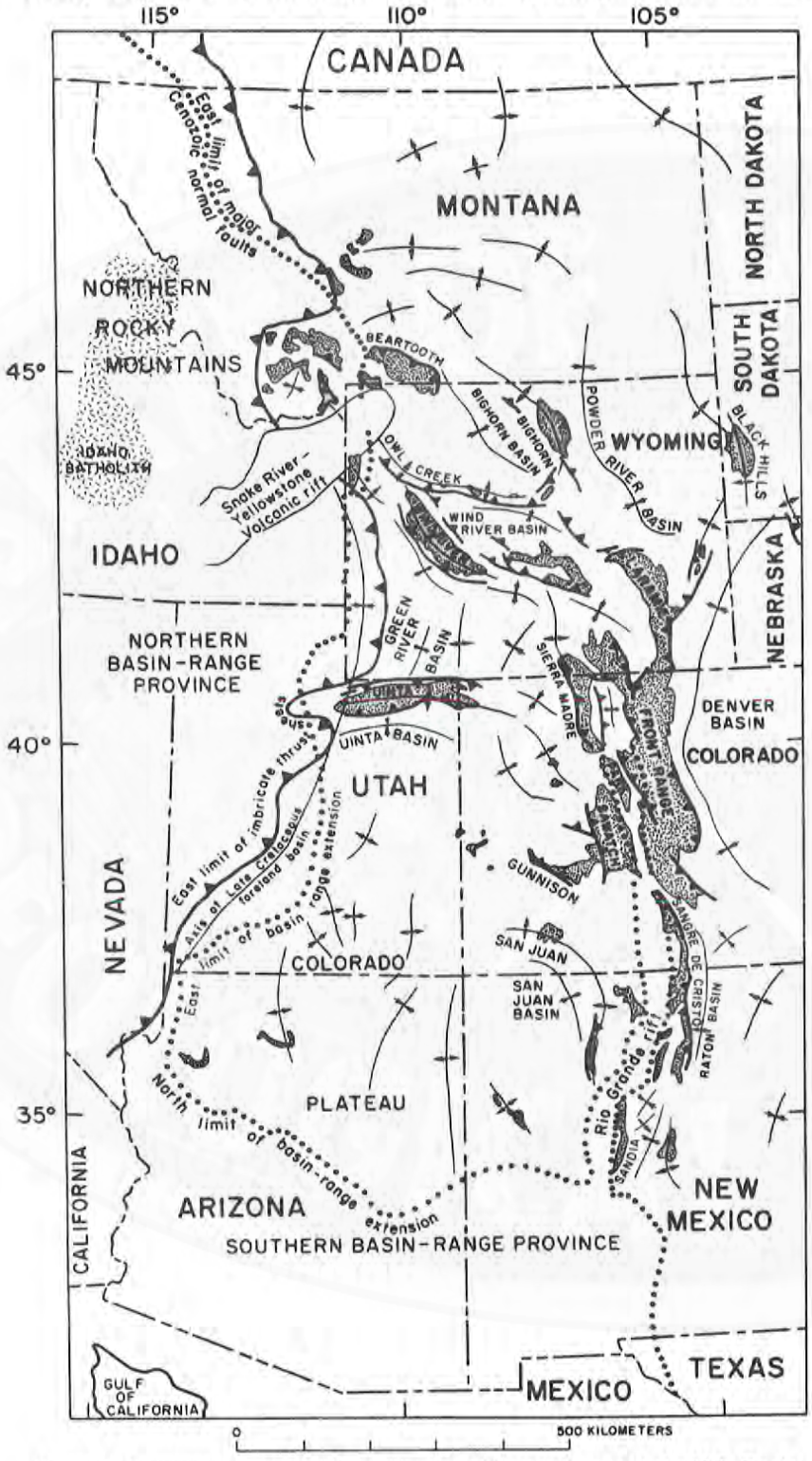


Figure 9. Basement-cored uplifts and basins of Laramide age in the U.S. Rocky Mountains. Precambrian basement rocks east of the foreland fold-and-thrust belt and north of the southern Basin and Range province are stippled [from Hamilton (1988b), figure 1; used with permission].

The term "Laramide" apparently was first used by J.D. Dana (1895, p. 359) to refer to the Rocky Mountain system, which forms an eastern cordillera in western North America from Canada to Mexico (Tweto, 1975). The name was derived from the poorly defined "Laramie series," a lithostratigraphic term used in Dana's time in reference to coal-bearing, nonmarine strata (Hayden's "Lignitic series") that lie above fossiliferous marine Cretaceous rocks in the Rocky Mountain region (so-called "Montana group"). The series, in turn, is unconformably overlain by Eocene Wasatch Formation or equivalents. Subsequently, the U.S. Geological Survey restricted use of the term "Laramie Formation" to the Denver Basin and designated this formation as entirely Cretaceous in age (Tweto, 1975). Perhaps in part because of its ambiguous initial definition, the term "Laramide" has been used to refer to various orogenic events that occurred during the late Mesozoic and early Tertiary. An example of such usage can be found in a recent paper by Maxson and Tikoff

(1996) who stated: "We use the term Laramide orogeny to describe all of the Laramide-age (80 to 45 Ma) structural, metamorphic, and magmatic features, at the latitude of California to Washington." Lageson and Schmitt (1994) discussed the usage of "Sevier" and "Laramide" and suggested that the terms had "outlived their usefulness," although in final analysis they retained the usage of Armstrong (1968), whereby Laramide is used for the basement-involved deformation of the Rocky Mountain foreland [see figure 1 in Lageson and Schmitt (1994)]. I also use Laramide in a geographically restricted sense, limiting it to refer to orogenic events in the eastern Rocky Mountains that occurred during the Late Cretaceous through mid-Eocene (Cross, 1986; Dickinson and others, 1988). As noted previously, debate continues on specific chronologic boundaries.

Yet another use of "Laramide" is to employ the term to describe the distinctive basement-cored uplifts in the U.S. eastern Rocky Mountains. In this usage, "Laramide-style" deformation denotes thick-skinned, thrust and reverse faulting that has involved Precambrian basement rocks of the Rocky Mountain foreland. As documented by Brown (1993), dips and orientations of these faults are variable; and in many cases, Laramide uplifts are bounded by conjugate, bivergent fault systems that form wedge-like, blocks of crystalline rock mantled by a thin cover of sedimentary rocks. True-scale representations of this distinctive style, in contrast to the thin-skinned deformation of the fold-and-thrust belt, are clearly depicted in Transect C-1 (Blake and others, 1989) or in cross section G (Oldow and others, 1989b, plate 7). However, basement involvement is also characteristic of the western part of the Sevier orogenic belt as well as its hinterland (Lush and others, 1988; Snoke and Miller, 1988; Yonkee, 1992; Royse, 1993).

Basement-cored uplifts of the Rocky Mountain region are truly anomalous in comparison with typical foreland deformation along the Cordilleran structural front. For example, in Canada, the Precambrian basement slopes gently westward and forms an unfaulted, 2 to 3° west-dipping ramp beneath various faulted and folded supracrustal rocks (Bally and others, 1966; Price and Mountjoy, 1970). At the beginning of the Laramide orogeny (Late Cretaceous) in Wyoming, the Precambrian-Paleozoic unconformity, east of the zero isopach of Late Precambrian sedimentary rocks, lay at -8000 to -10,000 feet (-2.4 to -3.0 km) (sea level) below a Paleozoic and Mesozoic sedimentary sequence (Stearns, 1971). A structure contour map of the present top of Precambrian basement indicates variations ranging from a restored, pre-erosion level of over 15,000 feet (4.6 km) above sea level in the Wind River Range to a level more than 30,000 feet (9.2 km) below sea level in the immediately adjacent northern Green River Basin (Blackstone, 1993b). This staggering structural relief was achieved chiefly by displacement along the Laramide, basement-rooted Wind River thrust in the early Tertiary (see Steidtmann and Middleton, 1991, for a detailed chronology of this faulting).

Structural style

The eastern Rocky Mountains, including the Wyoming foreland, is the "type" example of the uplift of basement rocks within a craton marginal to a thin-skinned, external zone of an orogen [see Rodgers (1987) for a worldwide summary]. Furthermore, fault zones associated with these basement-involved uplifts provide unparalleled opportunities to study the behavior of basement rocks during upper-crustal deformation.

Controversy and debate has surrounded structural interpretation of the Rocky Mountain basement-involved foreland uplifts, particularly as applied to subsurface geometries of the faults and deformational styles of basement rocks associated with the uplifts. Brown (1988, 1993) has traced the history of thought and controversy concerning these structures, so my comments are brief and focus on the most recent debate concerning vertical uplift models versus horizontal contractile models. The concept that Rocky Mountain basement-involved uplifts were principally vertical uplifts that occurred along steeply dipping faults evolved gradually, but probably is best summarized in a series of papers by David W. Stearns and coworkers (Stearns, 1971, 1978; Stearns and Weinberg, 1975; Stearns and Stearns, 1978). Their conceptual bases involve the following:

1. Basement-rooted fault zones are steep to subvertical;

2. Basement rocks, considered relatively homogeneous mechanically as contrasted to the overlying sedimentary cover, deformed by brittle fracture and did not fold; and

3. Uplift in the basement was partially compensated in the sedimentary cover as "forced folds," the so-called "drape folds" of Thom (1947).

During the 1970s, this school of thought dominated structural interpretations of basement-involved structures in the Rocky Mountain foreland. However, near the end of the 1970s, high-quality seismic reflection

studies (e.g., Smithson and others, 1978, 1979; Sharry and others, 1986) clearly demonstrated the actual thrust-fault geometry which, for example, involves substantial overhang of the Wind River thrust. This dramatic discovery underscored the importance of horizontal crustal shortening in the structural evolution of the Wyoming foreland (Kanter and others, 1981). Such shortening had been implied previously in the fold-thrust model of Berg (1962) as well as in some earlier cross sections (e.g., Anonymous, 1951). Subsequent systematic structural analysis of many of the Wyoming basement-involved uplifts by D.L. Blackstone, Jr., employing extensive use of well data, has clarified the importance of horizontal crustal shortening during the structural evolution of the Wyoming foreland (e.g., Blackstone, 1980, 1983, 1986, 1990, 1991).

Even though crustal shortening now has been demonstrated as a fundamental feature of the basement-involved structures in the Wyoming foreland, the role of basement deformation is still highly controversial and represents the focus of much recent research (e.g., Schmidt and others, 1993). A fundamental problem in this regard involves the following question: Does the basement actually fold during foreland shortening? This is not a new problem and has been discussed in respect to foreland tectonics for many years (e.g., Lees, 1952; Hudson, 1955). In a short essay on this subject, Matthews (1986, p. 5) concluded "... that structural basement did not fold during the Laramide." He contended that a planar unfolded contact typically exists between basement and the overlying sedimentary rocks, wherever such a contact is well exposed. Furthermore, he argued that the basement experienced rigid-body rotation during uplift (rather than folding), an interpretation that Erslev (1986) favored in order to balance the basement in cross sections of foreland uplifts.

Recently, the focus of this debate has changed. The question is not if the basement rocks actually fold, but rather what is the magnitude of folding (E.A. Erslev, personal communication, 1991). In his trishear model, Erslev (1991) argued that the style of folding in the sedimentary cover rocks (e.g., **Figure 10**) would not have to be mimicked by underlying basement rocks if this folding had been focussed downward into a



Figure 10. Aerial view of the plunging Beer Mug anticline, Freezeout Hills, Carbon County, Wyoming. The exposed core of the fold is Pennsylvanian Tensleep Sandstone, whereas the Permian Goose Egg Formation is exposed on the west (i.e., left) flank of the fold. Ridge-former horizons in the Goose Egg Formation are limestone-rich members. The fold is separated from the low, rolling area in front of the fold by the oblique-slip Indian Spring reverse fault (Taylor, 1996; Lillegraven and Snoke, 1996). A synformal anticline is defined by the Mesozoic stratigraphy that forms the footwall of the structure. Photograph by R.C. Thurston, Department of Botany, The University of Wyoming.

basement-rooted, brittle shear zone. Folding in the sedimentary cover therefore converges downward into a strain-compatible, propagating fault, and the entire system is modeled as a triangular shear zone in profile [i.e., Erslev's (1991) "trishear" model for fault-propagation folding]. Such fault-propagation fold models for Laramide basement-cored uplifts provide new approaches, which allow construction of balanced cross sections that account for both the uplift of the Precambrian crystalline basement and the deformation of sedimentary cover rocks. It is interesting to note that Blackstone's (1940) interpretative cross sections of the Pryor Mountains, Montana, are startlingly similar. His sections represent early examples of modern thinking regarding evolution of Laramide basement-involved uplifts.

Synorogenic sedimentary rocks

A broad foreland basin east of an active fold-and-thrust belt occupied much of the Central Rocky Mountain region prior to initiation of the Laramide orogeny in the Late Cretaceous (Jordan, 1981). Although isopach studies across this basin indicate differential rates of subsidence, marine strata are laterally continuous for long distances. This depositional pattern was terminated with the onset of the Laramide orogeny (Love and others, 1963). As individual basement-cored mountain blocks evolved in response to Laramide crustal shortening, the Rocky Mountain foreland was transformed into a complex system of structurally separated nonmarine basins [e.g., see Dickinson and others (1988), Lillegraven and Ostresh (1988), and the numerous earlier papers cited in these review articles]. Consequently, throughout the eastern Rockies, marine rocks of the Late Cretaceous (e.g., Lewis Shale in Wyoming) are succeeded by continental deposits of latest Cretaceous age (e.g., Lance Formation). These nonmarine rocks, in turn, are overlain by similar, commonly coal-bearing deposits of Paleocene age (e.g., Fort Union Formation).

Correlation of these nonmarine units throughout the Great Plains and Rocky Mountain region has been controversial. A complex lithostratigraphic terminology has been developed to replace the simple broad subdivisions first recognized by the early geologic explorers of the region. This terminology, in part, reflects the variety of depositional environments established in the evolving Laramide basins during the latest Cretaceous and Paleocene.

The early Laramide synorogenic sedimentary rocks are commonly overlain by nonvolcanogenic, arkosic, Lower Eocene continental rocks (e.g., the Wind River Formation of Wasatchian age) that reflect the continual erosion of mountain ranges formed during the Laramide orogeny. These Laramide deposits may be coarse bouldery facies deposited near rapidly eroding Laramide mountain fronts or fine-grained varicolored facies, commonly red-banded (Van Houten, 1948), interpreted as deposits of warm, humid lowlands that existed farther out into the Laramide basins. The Wind River Formation and equivalents were locally deformed during late-stage displacement along Laramide contractile faults. This sedimentary succession can also be unconformably overlain by still later Laramide or post-Laramide deposits (e.g., Steidtmann and Middleton, 1991), indicating continued erosion of Laramide highlands and commonly the input of detritus from new source areas (e.g., Absaroka volcanic field). The structural and stratigraphic relationships of Lower Eocene rocks with regard to younger contractile features and superjacent strata therefore can provide important constraints on establishing the end of the Laramide orogeny.

Orogenic timing in the eastern Rocky Mountains

During the development of Cordilleran plate-tectonic models in the late 1970s and early 1980s, the bracket 80 to 40 Ma was applied commonly to the Laramide orogeny (e.g., Coney, 1978; Dickinson and Snyder, 1978; Cross and Pilger, 1978b; Hamilton, 1981). Recent structural (Cross, 1986) and stratigraphic (Dickinson and others, 1988) syntheses have refined the timing, as well as investigated interregional variations. An important conclusion that comes from these studies, especially from Dickinson and others (1988), is that the Laramide orogeny was diachronous in the eastern Rocky Mountains. Using a variety of criteria, Dickinson and others (1988) argued that the orogeny ended near the close of the Wasatchian (about 51 Ma) in Wyoming. In contrast, the end of the Laramide orogeny in the southern Rocky Mountains (Colorado and New Mexico) was distinctly younger, perhaps between 35 and 40 Ma.

The time of initiation of the Laramide orogeny has always been more controversial. Plate-tectonic models favor initiation at about 80 to 75 Ma based on patterns of igneous activity and variations in convergence rates between the North American and eastern Pacific plates (e.g., Coney, 1978). According to Dickinson and others (1988), the sedimentary record indicates that Laramide deformation began in the Maastrichtian (75 to 66 Ma) and suggests no systematic areal diachroneity. In contrast, recent studies by Perry and

others (1991) suggested a southeastward migration of Rocky Mountain foreland deformation beginning in southwestern Montana earlier than 80 Ma and reaching the Colorado Front Range in the late Maastrichtian. Furthermore, these authors argued for a younger, northeastward-advancing deformational front across Wyoming. The migration of these deformational fronts is inferred to be related to an advancing and laterally spreading deep crustal detachment (Perry and others, 1991).

A mid-Cretaceous or older initiation of the Laramide orogeny has been suggested by some authors (e.g., Steidtmann and Middleton, 1991) to explain various sedimentological patterns in rock units of the Western Interior basin (e.g., sandstone channels in the Muddy Sandstone). However, these subtle variations in sedimentation do not necessarily require significant tectonic uplift. Dickinson and others (1988) outlined a series of criteria useful in recognizing the initiation of the Laramide orogeny. Of the dozen specific points enumerated by Dickinson and others (1988), initial development of local depocenters defined by isopachs and recognition of locally derived clasts within the basin fill are especially valuable constraining clues. Consequently, the conclusion by Love and others (1963) that the Laramide orogeny must postdate the last major marine Cretaceous transgression (i.e., Upper Campanian-Lower Maastrichtian Lewis Shale) is still the most logical upper limit of the initiation of regional deformation.

Plate-tectonic setting

During the late Cretaceous through early Tertiary, virtually the entire western North American Cordillera was involved in a complex orogeny that included variable structural styles (**Figure 11**). In Alaska and northern Canada, the orogeny was manifested by large-scale strike-slip faults, which record hundreds of kilometers of dextral translation and associated transpressional deformation (Eisbacher, 1985). Farther south, in Canada and the U.S. Northern Rocky Mountains, thin-skinned thrusting formed an impressive fold-and-thrust belt east of the Rocky Mountain trench (Price and Mountjoy, 1970; Price, 1981). In the Central and Southern Rockies, the cratonic crust of the foreland was tectonically shortened by deep-rooted, thick-skinned thrust and reverse faults that uplifted the basement rocks as adjacent intermontane basins developed. In the Colorado Plateau, enormous monoclines developed along reactivated basement faults (Huntoon, 1990). In southern California, the base of the continental crust was truncated as quartzofeldspathic metasedimentary rocks and associated oceanic rocks (Pelona-Orocopia-Rand schists) were thrust beneath crystalline basement rocks (Jacobson, 1983; Hamilton, 1989). In Mexico, a late Mesozoic to early Eocene fold-and-thrust belt, developed chiefly in late Mesozoic carbonate and clastic rocks, forms the Sierra Madre Oriental (Suter, 1987; de Cserna, 1989).

In an interpretation based principally on regional geologic arguments, Lowell (1974) and Dickinson and Snyder (1978) favored shallow-dipping subduction to explain the classic Laramide orogeny of the Rocky Mountain region in the western U.S. This "unfamiliar mode" of subduction, as characterized by Dickinson and Snyder (1978), was compared to segments of the modern Andean orogen. There, the subduction mode is amagmatic, with a diffuse, subhorizontal seismic zone. This analogue was further developed by Jordan and Allmendinger (1986), who demonstrated that the Laramide-style, basement uplifts of the Sierras Pampeanas of Argentina are coincident with a segment of subhorizontal subduction of the Nazca plate.

During latest Cretaceous and early Paleogene time, the rate of plate convergence between the North American and eastern Pacific plates increased markedly (Coney, 1972, 1978; Jurdy, 1984; Engebretson and others, 1984) (**Figure 12**). This change in relative plate motions led some tectonicists to postulate that the rapid overriding of subducting Pacific lithosphere by the North American plate was the fundamental cause of the shallow-dipping subduction mode (e.g., Coney and Reynolds, 1977; Cross and Pilger, 1978b). However, the increased convergence rate between the North American and eastern Pacific plates was not regionally restricted. Rather, it probably characterized the entire western North American continental margin (Henderson and others, 1984). Therefore, the cause of the shallow-dipping subduction mode must be more complex. If the subducting oceanic lithosphere was young and hot (Engebretson and others, 1984) (**Figure 12**), a shallow-dipping or flat-slab subduction geometry could have resulted. The age of the Farallon plate's eastern edge became progressively younger throughout the Laramide orogenic interval (**Figure 12**). Furthermore, Engebretson and others (1984) suggested that the gradual slowing of Farallon-North American convergence during the mid-Tertiary was also related to increased buoyancy of the subducting oceanic crust.

Livaccari and others (1981) and Henderson and others (1984) also used arguments involving buoyancy to account for the flat-slab subduction mode, as well as for the restricted basement-involved crustal shortening of the Rocky Mountain region. These authors suggested the subduction of unusually thick oce-

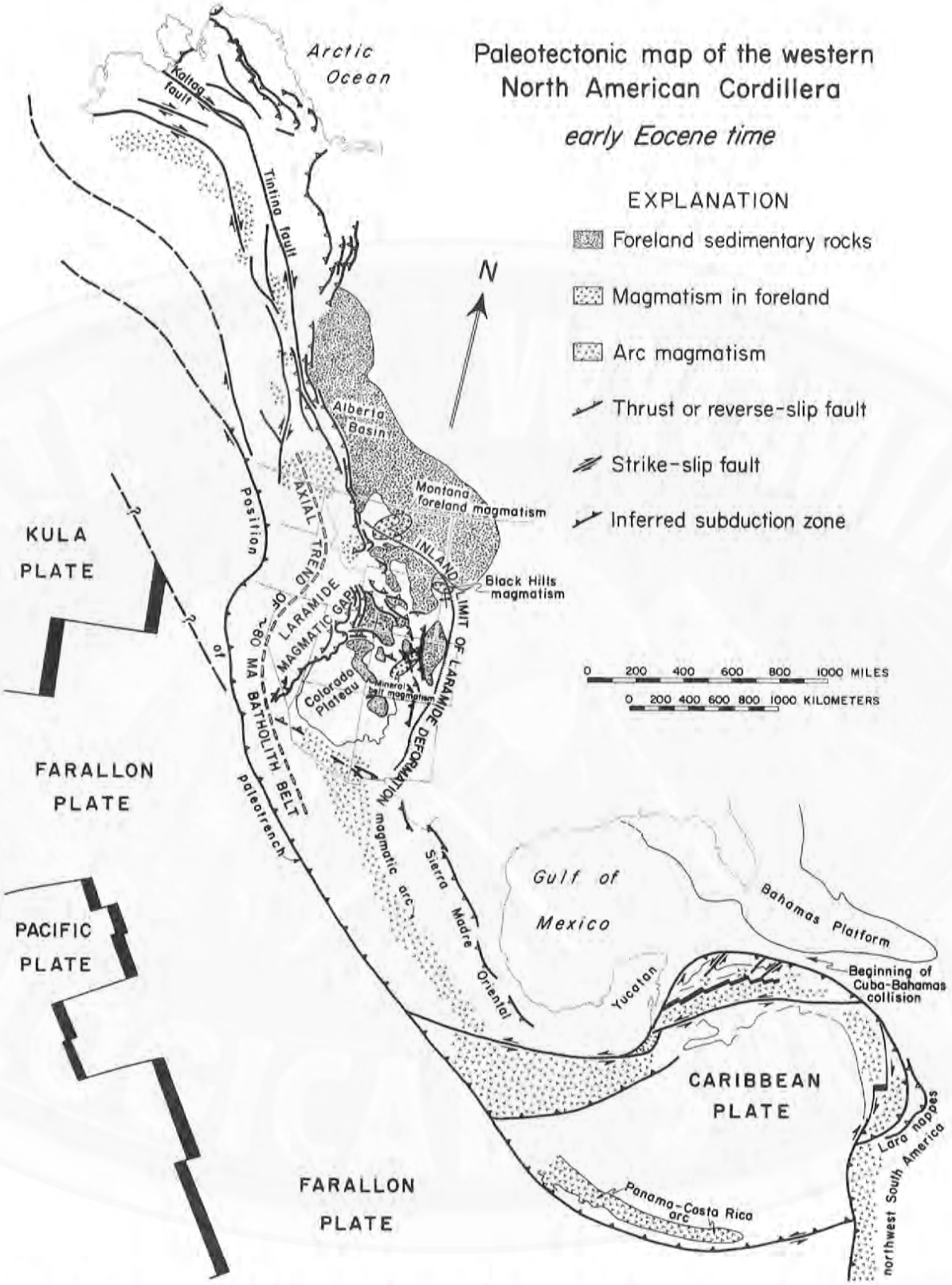


Figure 11. Generalized paleotectonic map of the western North American Cordillera at early Eocene time showing representative features that developed from Late Cretaceous through early Tertiary. Note that the shapes of some states have been modified [after Dickinson (1991), figures 10A, B] to account for Tertiary extension that occurred after the early Eocene. This map is a compilation of features derived from King and Edmonston (1972), Pindell and Barrett (1990, plate 12 of volume H), Dickinson (1991, figures 10A, B), Coney (1987, figure 3), and Engebretson and others (1985, figure 3d).

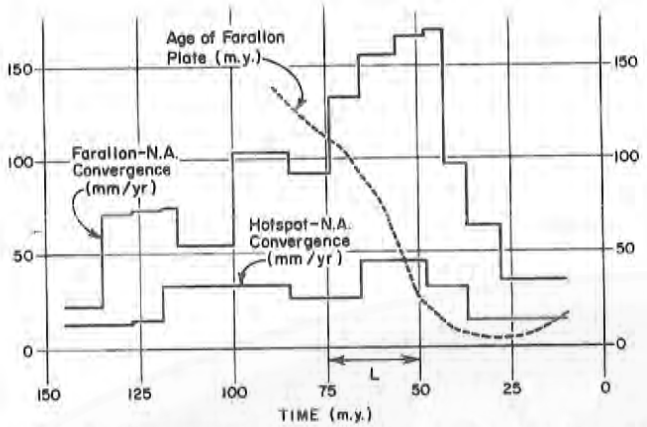


Figure 12. Components of motion of Farallon plate relative to North America and of North America over hotspots, perpendicular to an assumed coastline trend of N40°W near present San Francisco. Dashed curve is approximate age of Farallon plate that was subducted near the latitude of San Francisco (adapted from Engebretson and others, 1982). L denotes the Laramide orogeny, double-barbed arrow is the inferred length of orogenic activity in Wyoming (after Dickinson and others, 1988).

anic crust within the Farallon plate. Such anomalous crust could have been an aseismic ridge or oceanic plateau on the Farallon plate. If the feature was elongated, such as an aseismic ridge, its subduction, and consequently the locus of shallow subduction, would have migrated with time. Henderson and others (1984) suggested that an aseismic ridge, oriented north-northeast, was subducted during the Laramide orogeny, thereby causing the well-documented, southward migration of the Laramide magmatic null or gap. Their model also predicts a southward migration of inland deformation as recently suggested by Perry and others (1991) but, as yet, not demonstrated in detail.

A different, but fundamentally related, argument concerns the role of crustal thickening in the Rocky Mountains. Bird (1984, 1988) noted that estimates of upper crustal shortening across the Rocky Mountain foreland, typically about 5 to 15% (e.g., Kanter and others, 1981; Brown, 1988), are not sufficient to explain the magnitude of inferred crustal thickening that occurred during the Laramide orogeny. Bird (1984, 1988) postulated that this crustal thickening occurred by massive transport of deep crust and mantle lithosphere from southwest to northeast into the Rocky Mountain foreland, thereby substantially thickening the crust over a large region. This model requires a deep-crustal, high-strain zone (i.e., a zone of large-magnitude simple shear) that accommodated the massive transfer of lower crustal rocks beneath a huge area. Bird (1988) estimated that such a shear zone was 2 to 3 km thick, and recorded shear strains on the order of 1000. The shear tractions developed along this deep-crustal shear zone were transferred to the upper crust by basement-rooted faults that characterize the Rocky Mountain foreland structural style.

A fundamental problem with Bird's model (1984, 1988) is that geophysical studies in southeastern Wyoming (e.g., Johnson and others, 1984) indicate that variation in crustal thickness is directly related to the boundary between Precambrian provinces (i.e., the Cheyenne belt). Seismic refraction studies by the U.S. Geological Survey, show that crust beneath the Wyoming province is substantially thinner (37 to 41 km) than crust beneath the Colorado province (48 to 54 km). In a review of crustal structure of the Rocky Mountains, Prodehl and Lipman (1989) generalized variations in crustal thickness to conclude that it is thickest in southern Montana and Colorado and is 10 to 15 km thinner in the Central Rocky Mountains (i.e., chiefly Wyoming). These observations are difficult to reconcile with massive transport of lower crustal rocks from southwest to northeast during the Laramide orogeny, as postulated by Bird (1984, 1988).

A collisional plate-tectonic model for the Laramide orogeny has been proposed by Maxson and Tikoff (1996). These authors argue that a large, composite terrane (Baja British Columbia), presently situated in western British Columbia and southeastern Alaska, acted as a collider during Laramide time but subsequently was translated northward along a hypothetical, dextral transpressional fault system on its eastern margin. This transpressive collision would therefore account for such diverse features as the block uplifts of the Rocky Mountain foreland, Sevier fold-and-thrust belt, and hinterland metamorphic welt (Maxson and Tikoff, 1996). An excellent summary of the Baja British Columbia hypothesis has been provided by Cowan and others (1997), and these authors propose a set of tests to evaluate the hypothesis. Further evaluation of this hypothesis in regard to the evolution of the Rocky Mountains is thus premature until the whole concept of Baja British Columbia is better evaluated by geologists working along the western margin of North America.

The role of the Colorado Plateau in the Laramide orogeny has been a subject of much speculation. On virtually any small-scale geologic map of the Rocky Mountain region (e.g., **Figure 9**), the Laramide basement-cored uplifts appear to wrap partially around the Colorado Plateau along its eastern and northern margins (Hamilton, 1981, 1988b). This geographic and geometric relationship led Hamilton (1981, 1988b) to suggest that rotation of the Colorado Plateau was a key to basement-involved shortening of the Rocky

Mountain foreland. His model involves the clockwise rotation of the plateau simultaneous with flat-slab subduction. He speculated that the Euler pole of rotation was near central New Mexico, and the rotation was about 4° clockwise (Hamilton, 1988b). Detailed comparative paleomagnetic data from correlative upper Paleozoic and Triassic strata from the Colorado Plateau and craton support a clockwise rotation of 11°+/-4° for the plateau relative to the craton (Steiner, 1986). Other investigators (Bryan and Gordon, 1990), using a larger but perhaps less reliable data set, have also found evidence of systematic differences between plateau poles and stable North America, but favored a more modest clockwise rotation of 5°+2.4°/-2.3°. In either case, paleomagnetic data support a clockwise rotation of the Colorado Plateau, as originally hypothesized by Hamilton (1981).

Another aspect of the Laramide orogeny in regard to plate-tectonic models is the orientation and possible variation of the maximum shortening direction during the Late Cretaceous to early Tertiary structural evolution of the Rocky Mountain foreland. Based on Coney's (1978) vector summations of relative motion of the North American and Farallon plates, Brown (1988) assumed a consistent convergence vector between the plates as oriented N40°E for the Laramide orogeny. However, a previous analysis of structural development and timing by Gries (1983) had argued for a systematic counterclockwise rotation of the direction of maximum shortening during the Laramide orogeny, as manifested generally in the Rocky Mountain foreland. In contrast, the study of the evolution of Laramide basins by Dickinson and others (1988) suggested the broadly synchronous initiation of the complex pattern of Laramide uplifts and basins. Likewise, Cross' (1986) general inventory of dated structures in the Rocky Mountain foreland also did not support the chronology of structural development proposed by Gries (1983). Nonetheless, systematic overprinting relationships that developed during the Laramide structural evolution are well documented in the Wyoming foreland (e.g., Wise, 1983; Wise and Obi, 1992; Bergh and Snoke, 1992), suggesting that a complex, perhaps heterogeneous, strain field may be an important aspect of this history. Finally, Erslev (1993) noted a difference in inferred stress orientations for the Sevier fold-and-thrust belt and central Laramide foreland uplifts and postulated multilevel detachment during Late Cretaceous–early Tertiary contraction. He argued that the fold-and-thrust belt parallels the orientation of the actual plate margin, whereas the orientation of many structural features of the Laramide foreland reflects the plate convergence vector (i.e., northeastward).

In summary, numerous plate-tectonic models and variants of these models have been proposed to explain the anomalous intraplate deformation of the Rocky Mountain foreland during the Laramide orogeny. The concept of shallow-dipping subduction has been the centerpiece for many of the plate-tectonic models; an important exception is the collisional model of Maxson and Tikoff (1996). Many important problems remain unresolved concerning the timing and nature of progessive deformation, variation in maximum shortening direction, importance of regional deep-crustal detachment, and role of the Colorado Plateau in foreland shortening.

Paleogene magmatism

Regional overview

About 80 million years ago, significant changes in the pattern of arc magmatism developed in the U.S. Cordillera (Armstrong, 1974; Snyder and others, 1976). During this episode of "Laramide" magmatism (about 80 to 55 Ma), an eastward sweep of magmatic activity is well documented from California across southern Arizona to New Mexico (Coney and Reynolds, 1977). In the Northern Rockies, arc magmatism advanced from the Idaho batholith into southwestern Montana (e.g., Boulder batholith and associated volcanic rocks). A zone of northeastward-advancing magmatic activity along the Colorado mineral belt dates between 72 and 60 Ma (Mutscher and others, 1987). Isolated magmatism occurred in the Black Hills between 62 and 50 Ma (DeWitt and others, 1986; Lisenbee and DeWitt, 1993) and at various localities in central Montana during approximately the same time interval (Hearn, 1989). A distinct magmatic gap developed over a large region including parts of California, Nevada, and Utah (Armstrong, 1974).

The 225 to 80 Ma phase of Cordilleran magmatic activity is commonly attributed to effects of steep-dipping subduction, the "familiar mode" of Dickinson and Snyder (1978). In contrast, the overall sweep inboard of magmatic activity during the Late Cretaceous to early Tertiary (80 to 55 Ma) and development of a magmatic lull has been explained as caused by flat-slab subduction beneath a broad area of the U.S. Cordillera (Dickinson and Snyder, 1978). Evolution of the thin-skinned Sevier fold-and-thrust belt overlapped both this magmatic transition and the crustal shortening associated with the development of Laramide basement-cored uplifts.

At approximately 55 Ma (near the beginning of the Eocene), this complex Mesozoic-early Cenozoic magmatic activity terminated, as did the shortening associated with A-type subduction, in both the Northern and Central Rocky Mountains. However, a major new magmatic and tectonic episode was initiated over a broad region of the Rocky Mountains north of present latitude 42°N during the mid-Eocene (Armstrong, 1978). The igneous rocks of this magmatic episode are particularly well developed in the Absaroka-Gallatin (Wyoming-Montana), Challis (Idaho), and Kamloops (British Columbia) volcanic regions. Furthermore, broad-scale crustal extension developed in many regions, apparently genetically associated with this renewed magmatism. Evolution of metamorphic core complexes in southern British Columbia, northeastern Washington, and central Idaho was closely associated with this Eocene magmatism (e.g., Wernicke and others, 1987; Armstrong and Ward, 1991).

Paleogene magmatic suites

In Wyoming, the igneous rocks of the Black Hills, Rattlesnake Hills, and Absaroka volcanic province are isolated remnants of Paleogene magmatism (**Figure 13**). As a group, these magmatic centers include a great diversity of igneous rock types, ranging in composition from calc-alkalic to alkali-calcic to alkalic. How these various igneous rocks are related, if at all, is uncertain. Geochronometric data from these various magmatic suites indicate a considerable age span, extending from ca. 62 to 38 Ma. The Absaroka volcanic province and the Rattlesnake Hills are dominantly post-Laramide, whereas the Black Hills magmatism is more difficult to classify in this regard. Lisenbee and DeWitt (1993) concluded that the main phase of magmatism basically postdated initiation of Laramide uplift of the Black Hills (ca. 64 Ma), but these authors also noted that the uplift may have continued with a second pulse extending into the late Paleocene and early Eocene. Furthermore, the magmatism in the Black Hills is probably part of a regional event that extended to the northwest into central Montana (**Figure 13**). In this light, the Black Hills magmatism is best considered syntectonic with the Laramide orogeny.

Igneous rocks of the Black Hills occur as various shallow-level intrusions and minor extrusive phases along a west-northwest-striking zone that transects the South Dakota-Wyoming border (**Figures 13** and **14**). Lisenbee and DeWitt (1993) speculated that an ancient Precambrian zone of structural weakness (a strike-slip fault zone?) controlled distribution of these igneous rocks. This igneous activity has been bracketed between 50 and 62 Ma (DeWitt and others, 1986), although some younger dates have been reported (McDowell, 1971; Staatz, 1983). These rocks are chiefly alkaline, locally including alkali-calcic rocks; many rocks are iron rich. The rock types of this igneous suite vary from mafic (relatively minor) to silicic and range from silica oversaturated to highly undersaturated, although the extremes are related to extensive subsolidus alteration (Shearer, 1990). Some unusual rock types, such as carbonatite, occur locally.

The Rattlesnake Hills igneous complex consists of about 50 plugs, dikes, and various small masses of extrusive volcanic rock exposed in central Wyoming on the northeast flank of the Sweetwater arch, just north of the North Granite Mountains fault system (Hoch and Frost, 1993). The igneous rocks that constitute this complex include both alkaline (sodium-rich) and subalkaline rocks. Two samples from this suite have yielded K-Ar feldspar dates of about 44 Ma (Pekarek and others, 1974). Magma emplacement postdated local Laramide folding and was probably controlled by a reactivated zone of crustal weakness in Archean basement. Various magmas were emplaced high into the crust in this area, and magma mixing was an important process in the development of the diversity of igneous rocks in the Rattlesnake Hills (Hoch and Frost, 1993). Scarce clinopyroxene-mica cognate inclusions (cumulates) also suggest the importance of crystal fractionation during the differentiation of this igneous suite. These rocks are petrographically and chemically most similar to the Eocene igneous rock suite exposed in the Bear Lodge Mountains area, northeastern Wyoming (Hoch and Frost, 1993).

The Absaroka volcanic province is exposed in the rugged Absaroka Range of northwest Wyoming (Sundell, 1993), with a northwestern extension into the Gallatin Range of southwestern Montana (Chadwick, 1969). This volcanic province covers an area of 10,000 square miles (2500 km^2) and constitutes a volume of approximately 7000 cubic miles (29,000 km^3) of volcanogenic rocks (King and Beikman, 1978). It is the largest, least-studied volcanic province in the conterminous western United States.

Chadwick (1970) recognized two northwest-trending, subparallel belts of eruptive centers in the Absaroka-Gallatin volcanic province: an older eastern Absaroka belt and a younger western Absaroka belt. Igneous rocks as old as 92 Ma have been recognized in the eastern Absaroka belt of Montana (Meen and Eggler, 1987), although the bulk of the Absaroka volcanic province is Paleogene. These belts perhaps reflect ancient zones of crustal weakness, which could date back to the Precambrian (e.g., the northwest-striking Precambrian faults of Schmidt and Garihan, 1986b).

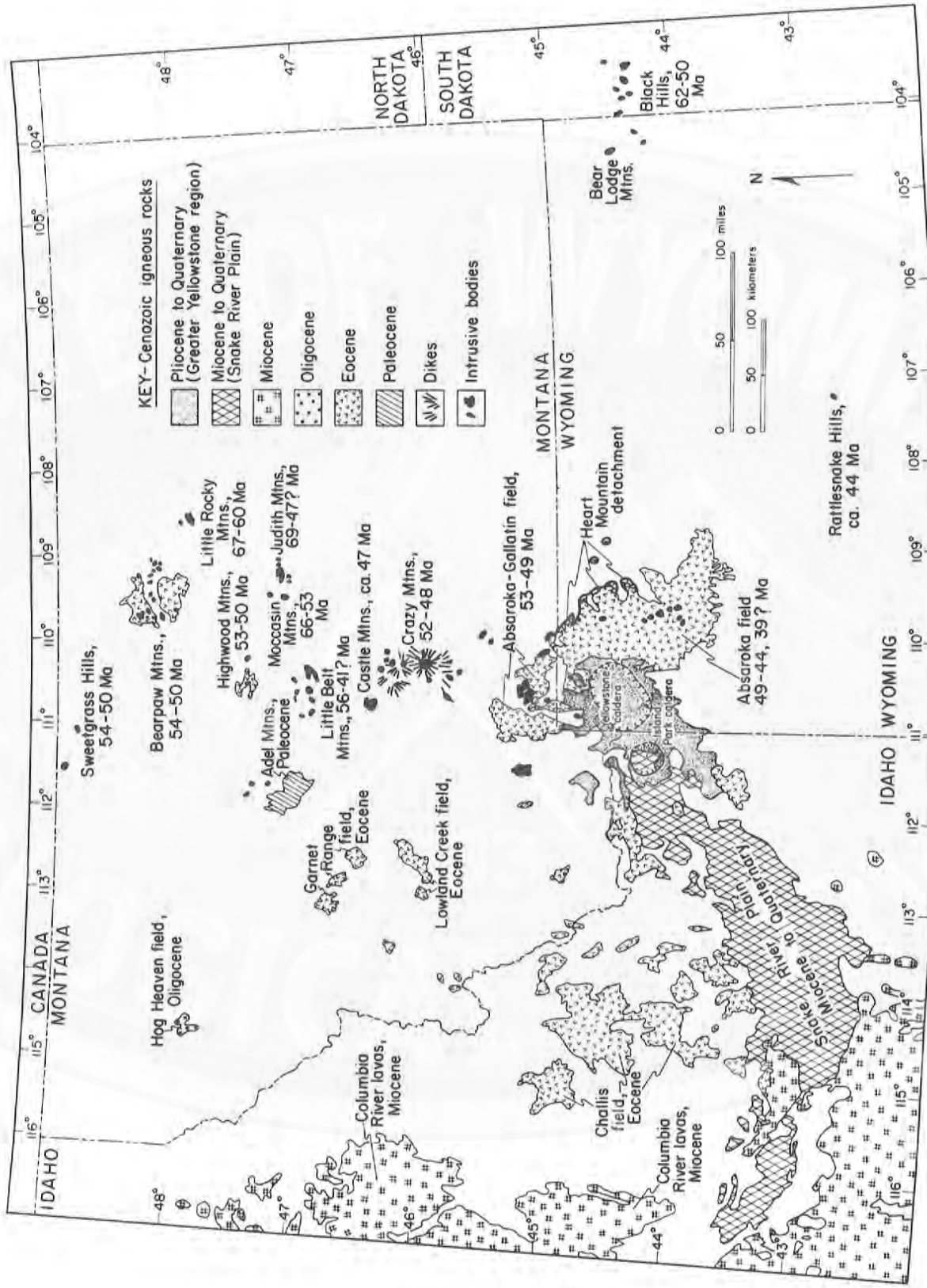


Figure 13. Cenozoic magmatism in the Northern Rocky Mountains, U.S. (after Chadwick, 1985, figure 2; King and Beikman, 1974; and Hearn, 1989, figure 1.1).



Figure 14. A panoramic view of Devils Tower-Missouri Buttes and environs, Black Hills uplift, northeasternmost Wyoming. Devils Tower rises about 1160 feet (~354 m) above the Belle Fourche River and reaches a maximum elevation of 5112 feet (1559 m) (U.S. Geological Survey Devils Tower 7.5-minute Quadrangle, Crook County, Wyoming, 1984). Devils Tower, considered an erosional remnant of a volcanic neck by many geologists, is an early Tertiary phonolite porphyry monolith characterized by spectacular columnar jointing. The Missouri Buttes exposed northwest of Devils Tower are smaller bodies of trachyte and phonolite, also with columnar jointing (Karner and Halvorson, 1987). These occurrences of igneous rocks in the northern Great Plains are examples of Laramide-age magmatism. Devils Tower was established as the first National Monument in 1906 by President Theodore Roosevelt. Photograph by the Wyoming Department of Transportation.

The pioneering geologists of Yellowstone Park and the northern Absaroka Range, under the direction of Arnold Hague, developed a six-part stratigraphic nomenclature for the Eocene volcanic rocks of the region (oldest to youngest): early acid breccia, early basic breccia, early basalt sheets, late acid breccia, late basic breccia, and late basalt sheets (Hague and others, 1896; Hague and others, 1899). Subsequent studies found various difficulties with this stratigraphic scheme, and gradually numerous other lithostratigraphic units were locally recognized and mapped. A modern reassessment of the volcanic stratigraphy of the Absaroka volcanic province was summarized by Smedes and Prostka (1972), who grouped all the Eocene volcanic rocks as the Absaroka Volcanic Supergroup. This supergroup is subdivided into three groups (oldest to youngest): the Washburn Group, Sunlight Group, and Thorofare Creek Group. In turn, each group contains various formations and members (Smedes and Prostka, 1972; Sundell, 1993). Coupled with the stratigraphic studies of Smedes and Prostka (1972) were new K-Ar radiometric age determinations [see figure 8, Smedes and Prostka (1972)] that indicated an age range from about 49 to 44 Ma for the Absaroka Volcanic Supergroup. An older age range from about 53 to 49 Ma has been reported in the Gallatin Range, southwestern Montana, for rocks of the Absaroka Volcanic Supergroup (Chadwick, 1970). In contrast, L.L. Love and others (1976) reported a fission-track sphene age of 38.8 ± 1.6 Ma from the southernmost plug (Washakie Needles) of the western Absaroka belt. The sum of these various age determinations perhaps suggests an overall pattern of progressively younger volcanism transgressing from northwest to southeast in the Absaroka volcanic province [L.L. Love and others (1976); see also Armstrong and Ward (1991), plate 1].

The Absaroka volcanic rocks are chiefly calc-alkalic andesitic and dacitic extrusive rocks, with subordinate potassic alkalic mafic lavas and rhyodacitic ash-flow tuffs (Smedes and Prostka, 1972). Small intrusive bodies are associated with these extrusive volcanic rocks. Complex facies relationships exist among various lithologies of the Absaroka Volcanic Supergroup, and two basic facies have been recognized: vent facies and alluvial facies deposits. The vent facies deposits are composed of autoclastic flow breccias, lava flows, mudflows, avalanche debris, and tuff. Stocks, plugs, laccoliths, and radial dike swarms also are common features of the vent facies. The alluvial facies deposits consist of aprons of well-bedded, reworked volcanic sedimentary rocks (see Smedes and Prostka, 1972, figures 6 and 7).

The Absaroka volcanic field is a series of deeply dissected Eocene stratovolcanoes that provide a remarkable window into evolution of the upper parts of a continental magmatic arc. In this light, it is important to emphasize that the Absaroka volcanic field was deposited unconformably across a complex paleotopography at the end of the Laramide orogeny (e.g., see Rouse, 1937). The subjacent rocks range in age from Precambrian to Tertiary, and some were strongly deformed during the Laramide orogeny (e.g., Love, 1939). In contrast, the volcanogenic rocks of the Absaroka volcanic province are only weakly deformed, if at all (Sundell, 1993).

An exception to the generally undeformed nature of the Absaroka volcanic rocks is the evolution of the Heart Mountain detachment fault, an enigmatic structure of virtually continuous controversy since its discovery [see Hauge (1993) for summary of the evolution of ideas and concepts]. A key aspect of the present controversy regarding the evolution of the Heart Mountain detachment fault concerns the involvement of Eocene volcanic rocks during faulting. One model–long advocated by W.G. Pierce (e.g., 1963, 1973, 1987)–suggests that only older Absaroka volcanic rocks were involved in the faulting and that younger volcanic rocks were deposited on a denuded detachment surface in spaces between distended blocks of Paleozoic rocks. The "continuous-allochthon" model of Hauge (1985, 1990), however, argues that all the Eocene volcanic rocks associated with the Heart Mountain detachment fault are

allochthonous and were part of an enormous, continuous allochthon that spread southeastward from a "break-away fault" (Pierce, 1960, 1980) in the northern Absaroka Range during the middle Eocene (Bridgerian). As outlined by Hauge (1993), this controversy has yet to be resolved and continues to be a subject of ongoing research.

Petrotectonic setting of Paleogene magmatism within the Wyoming province craton

Petrogenesis of Paleogene igneous rocks within the Wyoming province craton is poorly understood, but it is the subject of several detailed petrochemical studies (e.g., Dudás, 1991; O'Brien and others, 1991). Dickinson and Snyder (1978) related magmatism in central Montana and the Black Hills region to flat-slab subduction, thereby representing some of the most inland arc magmatism recognized in the western Cordillera. If Dickinson and Snyder's (1978) interpretation is correct, these igneous rocks were intruded and erupted approximately 750 to 930 miles (1200 to 1500 km), respectively, from their coeval Paleogene trench [see Heller and others (1987), figure 5].

Some geochemical data, principally from central Montana, have been cited as evidence of a subduction-related, arc component within the Paleogene igneous suites (e.g., O'Brien and others, 1991). The initial stage of this petrogenetic model was the development of a metasomatized carapace of phlogopite- and amphibole-bearing peridotite above a low-angle subducted slab (i.e., Farallon Plate lithosphere) at a relatively shallow depth (**Figure 15A**). Subsequently, this metasomatized carapace of asthenospheric mantle was dragged to greater depths by induced asthenospheric flow, where it broke down, forming melts that interacted during ascent with a lithospheric mantle keel beneath the Wyoming province craton (**Figure 15A**).

In contrast, Eggler and others (1988) argued that the overall geochemical characteristics of the Late Cretaceous and Paleogene magmatism within the Wyoming province craton does not indicate arc magmatism; and consequently, subducted material has not contributed chemical components to the magmatism. These authors suggested that the geographically widespread magmatism of the Wyoming province craton is back-arc in character, probably related to partial melting of a postulated mantle keel during extension and thinning of lithosphere and induced flow of asthenosphere (**Figure 15B**). This keel presumably consisted largely of depleted peridotite, locally modified by metasomatic enrichment. Detailed arguments against a subduction-related origin for the Paleogene magmatism of the Wyoming province craton were also presented by Dudás (1991), who favored the heating of lithospheric mantle either by upwelling asthenosphere or during regional decompression. O'Brien and others (1991) questioned if sufficient heat would be generated during advection in the asthenosphere to melt mantle lithosphere or if sufficient decompression occurred to cause melting by lithospheric extension. The tectonic setting of Paleogene magmatism within the Wyoming province craton is therefore still both enigmatic and controversial, despite substantial new chemical, isotopic, and geochronometric data.

Green River Formation — A Middle Eocene lake system

During the Middle Eocene, extensive lakes were developed in southwestern, central, and northern Wyoming [see e.g., Lillegraven and Ostresh (1988), figure 10] as well as in parts of Colorado and Utah. Love and others (1963) argued that the development of these lakes probably was related to a significant reversal in drainage, when the east-flowing drainage systems established in the Paleocene and early Eocene were cut off by an overall westward tilting of the Wyoming foreland. Synchronous with lake development in Wyoming was the initiation of extensive volcanism in the Absaroka Range, northwestern Wyoming, and in regions to the west (Chadwick, 1985). Lillegraven and Ostresh (1988) suggested that regional thermogenic upwarping may have been an important factor in the development of the extensive lake basin system.

Perhaps the best-studied ancient lake system in the North American Cordillera, or for that matter in the world, is Lake Gosiute, southwestern Wyoming. These ancient lake deposits have had a long history of scrutiny by a variety of geologists beginning with early explorations of the West [see Bradley (1964) for early references]. Their economic potential as the world's largest known trona reserve and as a source of oil shale have virtually guaranteed continuing analysis.

The Eocene lacustrine rocks are collectively referred to as the Green River Formation. These rocks form a gigantic lens between an enormous volume of fluviatile sediments divided into the Wasatch Formation below and Bridger Formation above (Bradley, 1964). Green River rocks are fine grained and gener-

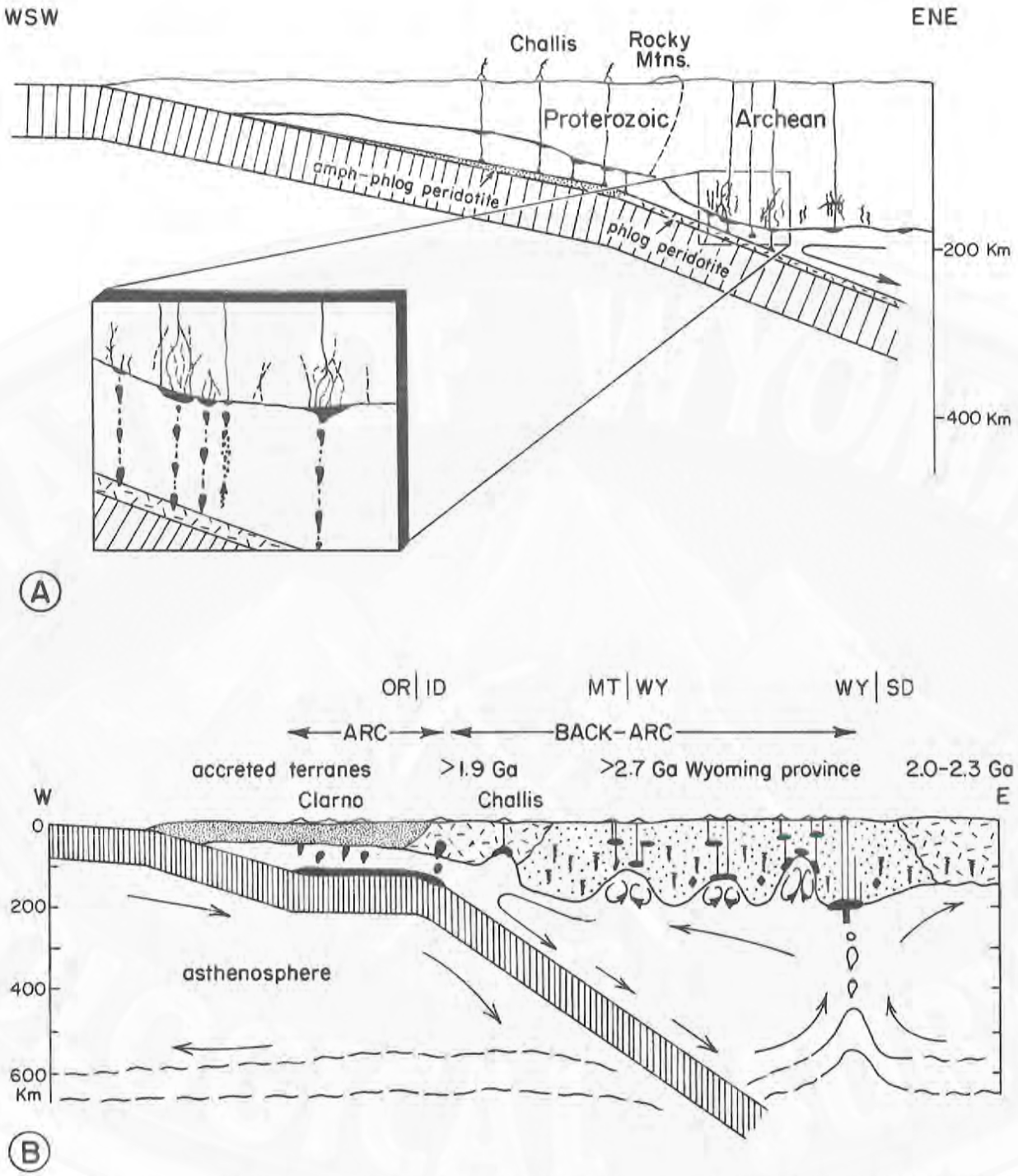


Figure 15. Contrasting petrogenetic models for Paleogene magmatism within the Wyoming province: (A) subduction-related arc magmatism associated with partial melting of asthenospheric mantle triggered by the infiltration of melts released from a metasomatized carapace formed above the low-angle slab of Farallon plate lithosphere (after O'Brien and others, 1991); and (B) back-arc magmatism associated with asthenospheric upwelling (after Eggler and others, 1988). Some details and criticisms of these models are discussed in the text.

ally calcareous. Rocks of both the Wasatch and Bridger formations are chiefly sandy mudstones with interbeds or lenses of sandstone; however, the Bridger Formation contains much more volcanogenic debris.

The history of Lake Gosiute is reflected in rocks of the Green River Formation [see Bradley (1964) for detailed stratigraphic descriptions]. The lake existed for about 4 to 8 million years and was characterized by

two main high stands separated by a major constriction (Surdam and Stanley, 1979). These various stages in the evolution of Lake Gosiute are recorded in the Tipton, Wilkins Peak, and Laney members of the Green River Formation. The classic interpretation of Lake Gosiute as an open-drainage, chemically and thermally stratified, meromictic lake was developed in a series of papers by W.H. Bradley (e.g., Bradley, 1948, 1964; Bradley and Eugster, 1969). In contrast, R.C. Surdam, H.P. Eugster, and coworkers argued that the depositional environment of the Green River Formation was, until late in its history, a shallow, closed-basin lake surrounded by broad playa flats (e.g., Eugster and Surdam, 1973; Surdam and Wolfbauer, 1975; Surdam and Stanley, 1979).

The Wilkins Peak Member of the Green River Formation is a distinctive unit in this ancient lake complex because it contains abundant saline minerals. These beds constitute an important economic resource, especially for trona ($Na_2CO_3 \cdot NaHCO_3 \cdot 2H_20$). Bradley (1959, 1964) recognized the significance of the Wilkins Peak Member in the depositional history of Lake Gosiute and reasoned that the lake had dramatically shrunk to about one-third the area of its maximum extent and had no outlet during this stage. In a detailed analysis of the Wilkins Peak Member, Eugster and Hardie (1975) described at least 50 major alternations of wet and dry periods during its deposition. When the climate was wet, the lake expanded and oil shale accumulated. During arid times, the lake shrank, and trona and halite precipitated in its central depths. Furthermore, Eugster and Hardie (1975) argued that flat-pebble conglomerates and mudcracks, characteristic of some rocks in the Wilkins Peak Member, indicate the importance of local desiccation. They envisioned extensive mud flats surrounding a shallow lake; the desiccated mud flats repeatedly were transgressed by the lake during wetter periods. Although deposition of the Wilkins Peak Member represents the most evaporative conditions in the history of the Green River Formation, the overall depositional framework outlined by Eugster and Hardie (1975, figure 19) for the Wilkins Peak Member is perhaps broadly applicable to the Tipton Shale Member (Surdam and Wolfbauer, 1975) and Laney Shale Member (Surdam and Stanley, 1979) as well. However, during deposition of upper parts of the Laney Shale Member, the extent and depth of the lake were considerably greater (Bradley, 1964; Surdam and Stanley, 1979), and thereby a meromictic depositional environment could have played a more significant role (e.g., Boyer, 1982).

Lake Gosiute was part of a large middle Eocene lake system that was an important late Laramide geomorphic feature in the west-central United States. Another large lake in this system was ancient Lake Uinta, which during times of maximum expansion occupied most of the Uinta and Piceance Creek basins in northeastern Utah and northwestern Colorado, respectively. These lake basins were connected by transitory outlets that, during high-water stands, allowed Lake Gosiute to overflow into Lake Uinta around the eastern end of the Uinta Mountains uplift (Surdam and Stanley, 1980). Hansen (1985) noted that the fossil fish fauna from the Eocene lakes had Mississippi River affinities, and therefore suggested that the Continental Divide in Eocene time was west of the lakes. In contrast, Dickinson and others (1988) speculated that an Eocene river from Lake Uinta led westward through the extinct thrust belt, draining to the Pacific Ocean.

A flood of volcaniclastic debris

Uppermost Eocene and lower Oligocene clastic rocks (i.e., White River Formation), rich in airfall tephra but surprisingly poor in crystalline basement debris, are widespread in southeastern Wyoming and adjacent states (Blackstone, 1975; Seeland, 1985). Deposition of these rocks marked the end of a period of intense late Eocene erosion that affected most of Wyoming (Lillegraven, 1993). Deposition of the White River Formation thereby signaled initiation of a new depositional cycle that occurred in a period of relative tectonic stability during the late Paleogene and most of the Neogene. The oldest rocks in the White River Formation contain mammalian remains of the Chadronian North American Land Mammal Age, which correlates approximately with an age of about 37 Ma on the geochronologic time scale (Lillegraven, 1993).

Although various lithotypes comprise these Chadronian and younger volcaniclastic deposits, fine-grained, off-white siltstone and bentonitic claystone are most characteristic (Denson and Bergendahl, 1961). Abundant volcanic debris that inundated the Great Plains and adjacent mountains in the late Paleogene and through the early Neogene were derived from remote sources; no evidence for Oligocene magmatism is known in Wyoming. Andesitic volcaniclastic rocks exposed in the Absaroka Range, once considered early Oligocene (e.g., Love, 1939; Love and others, 1963), have been shown to be Eocene by radiometric dating techniques (Smedes and Prostka, 1972; J.D. Love and others, 1976). Therefore, probably much of the volcanic debris was derived from voluminous mid-Tertiary pyroclastic vol-

canic activity to the west, the so-called "ignimbrite flare-up" of the Great Basin.

Rocks of the White River Formation form extensive, low-dipping exposures and commonly rest with marked unconformity on older rocks. A late Eocene unconformity also is well developed in the Southern Rocky Mountains (e.g., Epis and Chapin, 1975), and is part of a widespread Cordilleran hiatus formed prior to extensive mid-Tertiary volcanism (Gresens, 1981). The depositional environment of these latest Eocene and Oligocene rocks was principally fluvial, and a complex east-flowing drainage network existed in the early Oligocene (Seeland, 1985). This paleodrainage system distributed an extensive blanket of tuffaceous sediments from central Wyoming to the High Plains of western Nebraska and into adjacent South Dakota and Colorado. Denson and Bergendahl (1961) demonstrated that the thickness of the White River Formation varies from about 300 feet in southeastern Wyoming to 850 feet in northwestern Nebraska. At the close of the Oligocene, only the upper 1000 to 4000 feet of the highest peaks in the major mountain ranges of the Wyoming foreland protruded above an extensive aggradational plain (Love, 1960).

An important aspect of the Oligocene depositional history is that strata of the White River Formation were deposited across a region of locally marked relief. Although post-Laramide erosion created a regional unconformity (Gresens, 1981), local relief at the end of the Eocene in the Wyoming foreland was as great, if not greater, than at present. In southern and central Wyoming, large late Eocene to early Oligocene paleovalleys apparently existed, because undeformed White River rocks extend into the Precambrian crystalline cores of the Medicine Bow Mountains, Laramie Mountains, and southeastern Wind River Range (Evanoff, 1990).

The depositional cycle that began in the Chadronian continued throughout much of the Neogene, and large thicknesses of Miocene rocks are preserved in some Wyoming basins [see Love (1960), table 1]. Nevertheless, just prior to the deposition of basal Neogene sequences, yet another major erosional event affected broad areas of Wyoming (Lillegraven, 1993). This erosional interval is marked by a disconformity that typically occurs near the base of Arikareean sedimentary rocks (Denson and Chisholm, 1971; Trimble, 1980).

The lithostratigraphic terminology of the Neogene rocks of Wyoming is complex and controversial (Flanagan and Montagne, 1993). The name Arikaree Formation was first used by Darton (1899) for a series of gray sands characterized by layers of dark-gray concretions and exposed in northwestern Nebraska. These rocks are overlain unconformably by the more heterogeneous and locally gravelly Ogallala Formation (Swinehart and others, 1985). These units were deposited during the period from early to late Miocene; a previously used Pliocene designation for the Ogallala Formation has been discontinued because of recent corrections to the geologic time scale as used in North America (e.g., Berggren and others, 1985). The names Arikaree and Ogallala at first were restricted in usage to areas east of the Laramie Mountains. However, Denson (1965) suggested extension of the Arikaree-Ogallala nomenclature into central Wyoming, where a widespread succession of fine-grained sandstone, tuffaceous siltstone, and shale with subordinate interbeds of conglomerate, tuff, and limestone had been designated the Split Rock and Moonstone formations by Love (1961). This nomenclatural controversy remains unresolved, leading to various designations on published geologic maps and unfortunate confusion to all users. In a summary of the Neogene geologic history of Wyoming, Flanagan and Montagne (1993) suggested that use of the names Arikaree and Ogallala formations be restricted to latest Oligocene and Miocene rocks east of the Laramie Mountains, in accord with their original geographic use.

In summary, the latest Paleogene and Neogene rocks of southeastern Wyoming are part of a great sheet of outwash that spread along the eastern slope of the Rocky Mountains chiefly during the Oligocene and Miocene [see King and Beikman (1978), figures 2, 3, and 4]. In Wyoming, this sedimentary sheet was deposited across a late Eocene landscape that developed during extensive post-Laramide erosion. An important regional change is manifested by the transition from the humid climate of the early Tertiary to the semiarid climate of the middle Tertiary, marked by changes in floras and faunas of the intermontane basins (e.g., Dorf, 1959; McGrew, 1971). This fundamental climactic change has been cited as a principal factor contributing to mid-Tertiary aggradation (Mackin, 1947). Synchronous with this transition in climate was the inundation of the region with volcanic tephra principally derived from remote sources to the west. The existing streams consequently steepened their gradients in response to the increased load but decreased discharge, and a broad aggradational plain gradually formed that stretched across the Rocky Mountains and into the High Plains [see Love and others (1963), figure 9]. However, in the last 5 million years, erosion has greatly exceeded deposition (Trimble, 1980), and only a scant record of Pliocene deposition is preserved in western Nebraska (Swinehart and others, 1985). The various stratigraphic units that constitute the Mi-

ocene part of this outwash sheet are invariably bounded by unconformities, with channeling of underlying deposits during deposition of the younger strata (Skinner and others, 1977; Swinehart and others, 1985). The preservation of significant thicknesses of Miocene strata westward within the Rocky Mountains (e.g., Love, 1961) suggests that this great sedimentary sheet probably was widespread throughout most basins, and even across some of the ranges, of the Wyoming foreland.

Neogene extension

Neogene normal faulting is manifested locally throughout Wyoming. An episode of late Neogene extension has been long recognized in the fold-and-thrust belt (Armstrong and Oriel, 1965). Many of the broad, linear north-south valleys in the thrust belt are half-grabens, bounded on the east by listric normal faults (**Figure 16**). Some of these faults have reactivated segments of the thrust faults. An example is the relationship between the Hoback normal fault and the Prospect thrust as indicated by surface geologic mapping and a seismic reflection profile (Royse and others, 1975). Detailed structural cross sections by Royse (1993) indicate that ramps in the thrust fault trajectory localized the surface traces of the normal faults.

The most impressive manifestation of Neogene normal faulting in Wyoming is the spectacular, west-tilted Teton Range bounded on its east flank by the Teton normal fault (Smith and others, 1993). The Teton fault extends for about 44 miles (70 km) along the base of the eastern slope of the Teton Range, and exhibits stratigraphic separation, estimated locally, of at least 30,000 feet (Love and Reed, 1971; Love and others, 1973). The Teton fault is considered to be active, and an extensive set of adjacent Quaternary scarps suggests seismicity in the recent past (Smith and others, 1993). Although the north-striking Teton fault intersects northwest-striking, Laramide contractile structures at a high angle, Lageson (1987, 1992) argued that the Neogene normal fault may have reactivated the ramp of the Cache Creek thrust and was also controlled by pre-existing Precambrian shear zones.

Late normal faulting, although widespread in the Wyoming foreland (e.g., Love, 1970), is poorly understood in regard to its tectonic setting and regional significance. Some of these faults may not be of Neogene age and perhaps do not even indicate regional extensional deformation. Rather, these normal faults may have developed during, or shortly after, the Laramide orogeny in association with complex space problems created during crustal shortening or the relaxation of regional compressive stresses [e.g., Wise, 1963; also see Brown (1993)]. Nevertheless, important normal-fault systems are well documented at several localities in the Wyoming foreland and suggest that an episode of Neogene regional crustal extension affected this area (Hall and Chase, 1989). Three of the more important normal-fault systems of the Wyoming foreland include: the Continental fault system of the southern Wind River Range, the South Granite Mountains fault system, and the Whalen and Wheatland fault system of southeastern Wyoming.



Figure 16. Aerial view of Afton, Wyoming (elevation = 6239 feet = 1903 m), in Star Valley with the Salt River Range (part of the western Wyoming fold-and-thrust belt) in the background. A range-bounding, Holocene normal fault forms the western boundary of the Salt River Range, whereas Star Valley is a half-graben. This late Cenozoic faulting is still active as indicated by Quaternary fault scarps and occasional earthquakes (Piety and others, 1992). Photograph by W.A. Reiners, Department of Botany, The University of Wyoming.

The Continental fault is a major normal-fault system that extends for at least 50 miles (80 km) along the southwest flank of the Wind River Range (Zeller and Stephens, 1969). It was originally delineated and named by Nace (1939). This fault system apparently occurs in the hanging wall of the Wind River thrust (Blackstone, 1991, 1993a). The Continental fault system consists of anastomosing normal faults, both sinuous and planar, that generally dip steeply to the north and northeast (Steidtmann and Middleton, 1991). There has been considerable discrepancy among reports of displacement along the Continental normal-fault system. Nace (1939) estimated a minimum throw of 1450 feet (442 m); Berg (1962) cited a throw of 250 to 1000 feet (76 to 305 m); and Love (1970) stated that the mountain block was downdropped about 2000 feet (610 m) along the Continental normal fault. Structural analysis by D.L. Blackstone, Jr. (1991, 1993a) suggests that this normal-fault system does not cut the plane of the Wind River thrust, and thereby caused only modest, late, normal-sense slip along the thrust. Steidtmann and Middleton (1991) suggested that displacement along the Continental normal fault system occurred about 13 Ma.

The South Granite Mountains fault system refers to various normal-fault strands that occur along the northern margins of the Seminoe, Ferris, and Green mountains (**Figure 17**). The fault system, which strikes approximately N70°W, consists of steep, northward-dipping normal faults. Love (1970) estimated a minimum displacement of 2000 feet (610 m) along the fault system. Sales (1983) referred to this fault system as a major foreland collapse zone. Cross sections by Blackstone (1991), which cross the South Granite Mountains fault system and the Emigrant Trail thrust fault, suggest that the amount of late normal-sense displacement along the South Granite Mountains fault system was only a small fraction (perhaps 7% of total displacement) of the reverse-sense displacement along the Emigrant Trail thrust that occurred during the Laramide orogeny.

The Whalen and Wheatland faults of southeastern Wyoming are examples of northeast-striking normal faults. The Whalen fault strikes N43°E and dips steeply (74 to 77°) to the southeast (McGrew, 1963). This fault cuts the Upper Oligocene-Lower Miocene Arikaree Formation and exhibits about 600 feet (183 m) of stratigraphic separation. The Wheatland fault occurs northwest of the Whalen fault, strikes subparallel to it, but dips to the northwest. The Wheatland exhibits about 650 feet (198 m) of vertical separation (McGrew, 1963). These faults do not appear to be related to Laramide thrust faults, but they do parallel the inferred strike of the Cheyenne belt, again suggesting the importance of tectonic heredity in the evolution of foreland normal-fault systems.

Although late Cenozoic normal faults are relatively common in the Wyoming foreland, typically they record only modest displacements and represent only a few percent extension. Commonly these normal faults appear to reactivate older structures. The exact regional tectonic regime responsible for this late Cenozoic extension is uncertain. Similar structural relationships involving reactivation of older Laramide thrusts have been well documented in southwestern Montana (Schmidt and Garihan, 1986a; Lageson, 1989), but in this region the extensional deformation is attributed to the overprinting of Basin and Range structure onto a part of the Rocky Mountain foreland (Reynolds, 1979). In the Southern Rocky Mountains, normal faulting also is widespread (e.g., Taylor, 1975), and many authors have suggested that the effects of the Rio Grande rift system extend at least to central Colorado (e.g., Tweto, 1979). The Wyoming foreland is therefore a region of anomalous Neogene crustal extension, located north of obvious effects of the Rio Grande rift system and east of the Basin and Range structural province. Neogene extensional history of the Wyoming foreland suggests a zone of tectonic transition situated between areas of significant crustal extension and the unextended northern Great Plains.

The great exhumation

The late Neogene paleogeographic map of Love and others (1963, figure 9) depicts Wyoming as a broad aggraded plain characterized by volcanism in the northwestern corner (Yellowstone area) and Leucite Hills (northeastern Rock Springs uplift, southwestern Wyoming). According to their reconstruction, only scattered knobs of exposed Precambrian basement rocks, surrounded by Neogene sedimentary rocks, existed across the state. If this picture accurately represents late Neogene topography, fundamental changes had to have occurred to provide the mountainous scenery that now characterizes Wyoming.

The cause of this great exhumation of Neogene and older sedimentary rocks from the basins of Wyoming is a matter of considerable debate. Some workers argue that the erosion is evidence for epeirogenic uplift of the Rocky Mountains (e.g., Trimble, 1980). Others suggest that the exhumation can be explained by climactic deterioration (Molnar and England, 1990), and that



Figure 17. The late Cenozoic Kortes normal fault (Blackstone, 1965; part of the South Granite Mountains fault system) bounds the north flank of the Seminoe Mountains, south-central Wyoming. The fault trace is visible as a subhorizontal line (upper right center of photograph) that can be traced across the width of the photograph. The North Platte River cuts directly across the Seminoe Mountains through Black Canyon (left center of photograph), and this physiography is evidence of the superposition of drainage, a feature characteristic of many large river systems in the Southern and Middle Rocky Mountains (Atwood and Atwood, 1938). The white-weathering, low-dipping strata exposed adjacent to the North Platte River (right center of photograph) are Miocene sedimentary rocks. Photograph by R.C. Thurston, Department of Botany, The University of Wyoming.

the region has been near its present elevation since the mid-Eocene (the end of the Laramide orogeny). Regardless of the exact role of tectonic uplift and climate, an enormous volume of sediment was stripped from the area now known as the eastern Rocky Mountains during the late Cenozoic. Much of this sediment was transported to the Gulf of Mexico, where thousands of feet of Pliocene and Pleistocene deposits accumulated (e.g., Woodbury and others, 1973).

If the great exhumation was driven by tectonic uplift, then the mantle lithosphere beneath the Rocky Mountains may have thinned significantly during the past 5 million years to compensate for the uplift (Suppe and others, 1975; Eaton, 1986). Results from seismic studies (e.g., Herrin and Taggart, 1968; Grand, 1987) are consistent with the presence of thin or hot mantle lithosphere under the western United States, but the spatial resolution of such studies is low. Nevertheless, analyses of lithospheric flexure (Sahagian, 1987; Angevine and Flanagan, 1987; Mitrovica and others, 1989) suggest that as much as one-half of the modern elevation of the Rocky Mountains is compensated by thick accumulations of late Cretaceous and early Tertiary sedimentary rocks (Cross and Pilger, 1978a; Cross, 1986) and Laramide crustal thickening.

To settle the debate about the relative roles of climate and tectonism in the great exhumation, reliable estimates of the absolute elevation history of the Earth's surface are needed. One approach may be to determine paleoaltitudes using estimates of paleotemperature that are derived from physiognomic analyses of fossil flora. Studies of this type (e.g., Wolfe and Schorn, 1989; Gregory and Chase, 1992) suggest that the mountains of Colorado reached their present elevations by the early Oligocene. Similar studies of paleoflora in Wyoming would be of value, but fundamental questions regarding the large corrections and poorly known parameters used in computing paleoaltitudes must be addressed.

High-level erosional surfaces

Many mountain ranges in Wyoming exhibit an extensive subsummit surface, above which peaks project and below which streams and rivers have cut valleys and gorges to depths of several thousand feet. Recognition of high-level erosional surfaces in the Rocky Mountain region goes back to early geologic explorers (e.g., A.R. Marvine of the Hayden Survey). An interesting summary of various ideas concerning the number of recognized surfaces, their inferred ages, and their origins in regard to the Colorado Front Range was presented by Bradley (1987). In this review, he emphasized that accurate dating is the key to understanding the significance of these erosion surfaces. In this light, the apparently best dated erosion surface in the Rocky Mountains is the so-called "late Eocene" surface of the Southern Rocky Mountains (Epis and Chapin, 1975). Epis and Chapin based the age designation on recognition that an ancient surface, similar to exhumed surfaces, is overlain by radiometrically dated Oligocene volcanic rocks. Some Colorado workers [as cited in Bradley (1987)] argue that the famous Sherman surface of southeastern Wyoming is of similar age, although no Oligocene volcanic rocks are deposited upon it. Knight (1953) also concluded that the subsummit erosion surface of the Medicine Bow Mountains [his Libby Flats surface; but called Medicine Bow surface by Steven (1956)] was of late Eocene age and represented post-Laramide, regional planation. In contrast, Steven (1956) noted that a major paleovalley in the Medicine Bow Mountains (Kings Canyon), filled chiefly with early Oligocene White River Formation and locally with Miocene sedimentary rocks, is truncated by the subsummit erosional surface (see his figure 2), suggesting that the age of this surface is late Tertiary. Knight (1953) had previously suggested that the paleovalleys were relicts preserved despite extensive late Eocene planation and subsequently filled with Oligocene and younger sedimentary deposits.

Interpretations of the age and developmental history of various high-level erosional surfaces in the Rocky Mountains are still very controversial after many years of study and numerous published reports (Mears, 1993). Nevertheless, many past and recent workers concur that significant relief existed in the late Eocene (e.g., Knight, 1974, his figure 4) despite extensive post-Laramide erosion (see also Love, 1941, for a clear statement of this interpretation as well as Evanoff, 1990). Furthermore, many erosional surfaces in Wyoming may be of late Tertiary age. Mears (1993) concludes his review of high-level erosional surfaces in Wyoming with this interpretation, citing the preservation in the Laramie Mountains of a planar surface of deposition, upon which the Upper Miocene Ogallala Formation was deposited. This erosional surface and the Miocene sedimentary rocks deposited upon it form the famous "Gangplank," which connects the late Miocene depositional surface of the High Plains with the Sherman erosional surface in the Laramie Mountains (Mackin, 1947; Moore, 1959). Mears also supports Mackin's (1947) hypothesis that these surfaces are pediments. According to this interpretation, the high regional slopes of the subsummit erosional surfaces are more compatible with the semiarid climate of the late Tertiary than the inferred humid climate of the late Eocene. Consequently, the subsummit pediment could have formed at a high elevation.

Yellowstone Volcanic Plateau

The Quaternary Yellowstone Volcanic Plateau is at the northeastern end of the eastern Snake River Plain, a volcano-tectonic feature in the U.S. Cordillera that developed across the Basin and Range province (**Figure 1**). The Yellowstone Plateau stands at about 8200 feet (2500 m), although flanked by mountains that rise to nearly 13,000 feet (4000 m). To the southwest along the eastern Snake River Plain, the terrain drops gradually to a region of lower relief at about 3300 to 5000 feet (1000 to 1500 m) elevation. One widely-cited hypothesis attributes the development of the Snake River Plain to the southwestern migration of the North America plate across a hotspot fixed in the mantle and presently beneath the Yellowstone region (e.g., Morgan, 1972; Smith and others, 1974; Eaton and others, 1975; Smith and Christiansen, 1980; Pierce and Morgan, 1992; Smith and Braile, 1993). The vector of motion of the North America plate, inferred from other methods, is parallel to the azimuth of the eastern Snake River Plain (Minster and Jordan, 1978), and Leeman (1982) suggested that the hotspot trace can be extended to the ca. 16 Ma McDermitt volcanic field in north-central Nevada [also see Pierce and Morgan (1992)]. However, Hamilton (1989) argued that magmatism in the eastern Snake River Plain was erratic, and an orderly progression of magmatism to the northeast, as predicted by the hotspot model, cannot be demonstrated in detail. Hamilton (1989) concluded that the eastern Snake River Plain is a wedge-shaped extensional rift that has propagated northeastward with time, resulting in volcanic activity.

Regardless of the exact origin of the Snake River Plain, the Yellowstone Plateau is the product of voluminous volcanic eruptions during the past 2.2 million years (e.g., Christiansen, 1984). Three cycles are recognized in the magmatic history of the Yellowstone Plateau with climactic eruptions at about 2.0, 1.3, and 0.6 Ma (Christiansen and Blank, 1972). The rhyolitic ash-flow sheets that formed during these pyroclastic eruptions constitute the major stratigraphic units of the volcanic plateau, and are called (oldest to youngest): Huckleberry Ridge, Mesa Falls, and Lava Creek Tuffs of the Yellowstone Group (Christiansen and Blank, 1972).

The three cycles evolved in a basically similar fashion. Each cycle began and ended with a protracted period of intermittent lava eruptions of rhyolite and subordinate basalt but climaxed with an explosive, caldera-forming pyroclastic eruption of rhyolitic ash flows (Christiansen, 1984). These voluminous ash flows spread out over hundreds of square miles, and volcanic debris from these eruptions rose high into the atmosphere and was carried for thousands of miles. Deposits of Yellowstone ash are widespread throughout the western U.S. and southern Canada (Izett and Wilcox, 1981).

The record of the third volcanic cycle is most complete and that cycle accounts for much of the volcanic landscape in present-day Yellowstone National Park (Christiansen and Blank, 1972). This cycle began about 1.2 Ma. Rhyolitic lava intermittently erupted from arcute fracture systems, gradually outlining a large area of the roof of the subjacent magma chamber. This faulted carapace subsequently collapsed during a climactic pyroclastic eruption 630,000 years ago, forming an enormous caldera (Christiansen, 1984). During the formation of the Yellowstone caldera, more than 1000 km^3 of Lava Creek Tuff were so rapidly erupted (in hours or days) that there is no evidence of erosion, aqueous reworking, or the deposition of extraneous materials within the tuff sequence (Christiansen, 1984). However, detailed stratigraphic data indicate that the tuff consists of two sheetlike deposits that erupted so close in time that they welded and crystallized as a single cooling unit. Furthermore, the caldera is actually two overlapping ringlike fracture systems that form a compound feature about 47 miles (75 km) long and 28 miles (45 km) across (Christiansen, 1984). Resurgence of magma into the subjacent chamber caused uplift of the caldera floor, forming two structural domes (Smith and Christiansen, 1980). During the past 630,000 years, sediments and postcaldera rhyolitic lavas have covered much of the compound caldera basin. Furthermore, during the past 150,000 years, rhyolitic lavas have covered the western resurgent dome and completely buried the western rim of the caldera (Christiansen, 1984). Although no eruptions have occurred in the Yellowstone region in the past 70,000 years, the vigorously active hydrothermal systems manifested as geyers, hot springs, and fumaroles reflect the presence of a major, still-molten magma body beneath a large part of the Yellowstone Plateau (Eaton and others, 1975; Smith and Braile, 1993).

Holocene seismic activity in Wyoming

Although seismic activity is not common throughout much of Wyoming (Case, 1986), the Intermountain seismic belt crosses westernmost Wyoming, where it is defined by a zone of numerous earthquake epicenters (Sbar and others, 1972; Smith and Sbar, 1974). Several active fault systems are part of this belt, including the Teton fault, which has been assigned a maximum credible earthquake (MCE) of magnitude 7.5 (Gilbert and others, 1983). The largest historic seismic event (surface-wave magnitude or M_S=7.5) of the Intermountain region occurred on August 17, 1959, at Hebgen Lake, Montana, located only a few miles west of the northwestern boundary of Yellowstone National Park. The shock was felt over an area of 600,000 square miles (Murphy and Brazee, 1964). Twenty-eight people were killed–the disastrous quake-caused Madison Canyon Slide accounted for all but two of these deaths (Hadley, 1964).

Several large fault scarps, and many smaller ones, formed during the Hebgen Lake earthquake. The most spectacular scarp-fissure system could be traced for 14 miles (22.5 km), defining the Red Canyon fault scarp (Witkind, 1964). This set of scarps and fissures appeared in surficial material along or near the trace of a Tertiary normal fault (Red Canyon fault). The scarp had an average height of 10 feet (3 m), but locally it increased to about 20 feet (~6 m) (Witkind, 1964). The close association of earthquake-related scarps and Tertiary normal faults with thrust faults suggests that the recent displacement was normal-sense along reactivated thrust faults (Myers and Hamilton, 1964; Doser, 1985).

A summary of the distribution of historic earthquake epicenters across Wyoming (Case, 1986) indicates only scattered events throughout much of the state as compared to the concentration of seismicity along the western margin of the state. Furthermore, many seismic events in Wyoming east of the Intermountain seismic belt do not correlate well with mapped traces of known faults. However, such sum-

maries of seismicity in Wyoming are based upon a historic record only compiled since 1871 (Case, 1986). The long-term frequency of seismic events throughout Wyoming is therefore very poorly known.

Concluding remarks

The geology of Wyoming includes many representative features of the more general tectonic evolution of the North American Cordillera. During the past twenty or more years, models and syntheses of the geological evolution of the Cordillera have been based principally upon the paradigm of plate tectonics. To understand the geologic evolution of Wyoming in this perspective has proven to be a formidable challenge. Although inferred Proterozoic and younger tectonic settings seem to have many actualistic examples (i.e., passive margins, magmatic arcs, accretionary complexes, and so forth), plate-tectonic models for the Archean are still controversial and probably oversimplified. Furthermore, the great distance of Wyoming from the plate margin since early Mesozoic time has impeded the correlation of events recorded at the western continental margin. Many future insights into the geological evolution of Wyoming will be based upon carefully designed studies that challenge the "conventional wisdom" espoused within this article.

In this light, I conclude my overview of the geology of Wyoming with a list of unsolved problems (chiefly tectonic in scope) that could evolve into important research topics:

1. Accretion history of the Archean Wyoming province;
2. Regional extent of deformation and metamorphism associated with Early Proterozoic accretion along the southern margin of the Wyoming province;
3. Tectonic setting of the Laramie Anorthosite Complex and Sherman batholith;
4. The role of the middle and lower crust during Laramide crustal shortening;
5. Cause(s) of regional variation in orientation of Laramide uplifts and basins;
6. Kinematic history of basement-cored uplifts and the role of an evolving or heterogeneous stress field;
7. Petrotectonic setting of Laramide and post-Laramide magmatism in the Rocky Mountain foreland;
8. Tertiary mean-elevation history;
9. Cause of crustal extension in central Wyoming; and
10. Cause(s) of late Cenozoic exhumation.

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Paleozoic history of Wyoming[1]

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Abstract

The Paleozoic Era began with Wyoming at the craton margin, above sea level, and near the equator. During the last half of the Cambrian Period, the sea advanced sporadically across the state, blanketing Precambrian rocks with sand and gravel. By the end of the period, when siliciclastic sediment finally covered the ancient crystalline rocks in northeastern Wyoming, the equivalent Cambrian facies in western Wyoming were buried under hundreds of feet of shale and limestone.

The state preserves little evidence of Early and Middle Ordovician deposition, but a Late Ordovician clear-water sea produced thick carbonate accumulations now represented by prominent cliffs of dolomite in northern and western Wyoming. Post-Ordovician erosion beveled the thick carbonate sequence from north to south, completely removing it from the southeastern area except for foundered blocks in small kimberlite diatremes near the Colorado border. Inclusions of Silurian marine fossils in a few diatremes prove that a sea covered the area sometime during the period.

By the time a Devonian sea invaded Wyoming, Silurian strata were lost to erosion and drainage systems had developed on older bedrock. Paleovalleys filled with estuarine strata record Early and Late Devonian transgressions. At the height of the Late Devonian transgression, marine deposition throughout the northern and western part of the state resulted in several hundred feet of Upper Devonian strata. Following a major regression, the sea returned to parts of Wyoming shortly before the Devonian Period ended and spread over most of the state in the ensuing period.

Wyoming's heritage from the warm, clear-water Mississippian sea is a wedge of carbonate strata thickening west and north from a zero edge in the southeast to more than 1000 feet at the western border. Carbonate production on the Wyoming shelf was halted by a major regression that moved the shoreline to the western border of the state, and a prominent karst topography developed. The overlying Amsden Formation includes Mississippian and Pennsylvanian marine strata; it has been the center of a long controversy over lateral and vertical relations of its members.

Wind played a major role in distributing a thick blanket of Pennsylvanian sand over the Wyoming shelf, although some of it was reworked during occasional incursions of shallow seas. In eastern Wyoming, marine conditions were more continuous and carbonate environments more frequent. Meanwhile, arkosic fluviatile sediments spread into southeastern Wyoming from the north end of the ancestral Front Range.

Subaerial erosion that affected the central part of the state during the Pennsylvanian-Permian transition expanded to include the marginal areas where deposition had continued into Early Permian time. The state-wide erosion surface was subjected to aggradation as several contrasting depositional environments migrated back and forth over the region. This resulted in complex intertonguing of redbeds (east), carbonates (central), sandstone (northwest), and phosphatic and cherty strata (west). The absence of a prominent disconformity between Permian strata and overlying Triassic mudstones contrasts sharply with paleontological evidence of missing record at the close of the era.

[1] This summary is a slightly revised and reformatted version of a paper originally published under the same title in *Geology of Wyoming*, Wyoming State Geological Survey Memoir 5 (1993), p. 164-187.

Introduction

Humans have been examining Paleozoic rocks of the area we now call Wyoming for at least 10,000 years. For almost 99% of that time, investigations were limited to work by a few prospectors seeking to maintain or improve their quality of life in terms of projectile points, pigments, cave sites, and spring locations. Although these people published nothing aside from a few petroglyphs, the immigrants who followed them have, in the last century, produced an immense written record of observations and ideas about the rocks of interest here. The evidence they have interpreted represents all seven Paleozoic periods, albeit by no means equally (**Figure 1**). The following synthesis will not satisfy those who have mastered and (or) generated any significant part of that record, but it should be of use to those preparing to join that group and to those who are mildly interested in the topic but don't want to take their coats off.

The title should be taken literally; this is not an article on Paleozoic stratigraphy of the state. Although many important formations are mentioned by name, no attempt is made to review the many complexities of nomenclature that reflect different preferences on the part of stratigraphers separated by generation, geographic location, and (or) philosophy. Likewise, little or no mention is made of many unresolved questions concerning correlation and detailed age assignment of individual stratigraphic units. Readers interested in these topics will find help in the bibliographies of the papers cited and from the correlation chart by Love and others (1993).

Although the references noted in this review represent only the tip of the iceberg, they provide entry to the wealth of literature on the Paleozoic geology of Wyoming. Some references not limited to a single geologic period can be mentioned at the beginning. The *Geologic map of Wyoming* (Love and Christiansen, 1985) is accompanied by a description of stratigraphic units (sheet 2) and an extensive list of references (sheet 3). The large atlas published by the Rocky Mountain Association of Geologists (Mallory, 1972) provides several thickness and lithofacies maps for each system. Of several general surveys of regional Paleozoic history, recent ones by Peterson and Smith (1986) and Peterson (1987) are especially useful for the citations of previous literature. An interpretation of

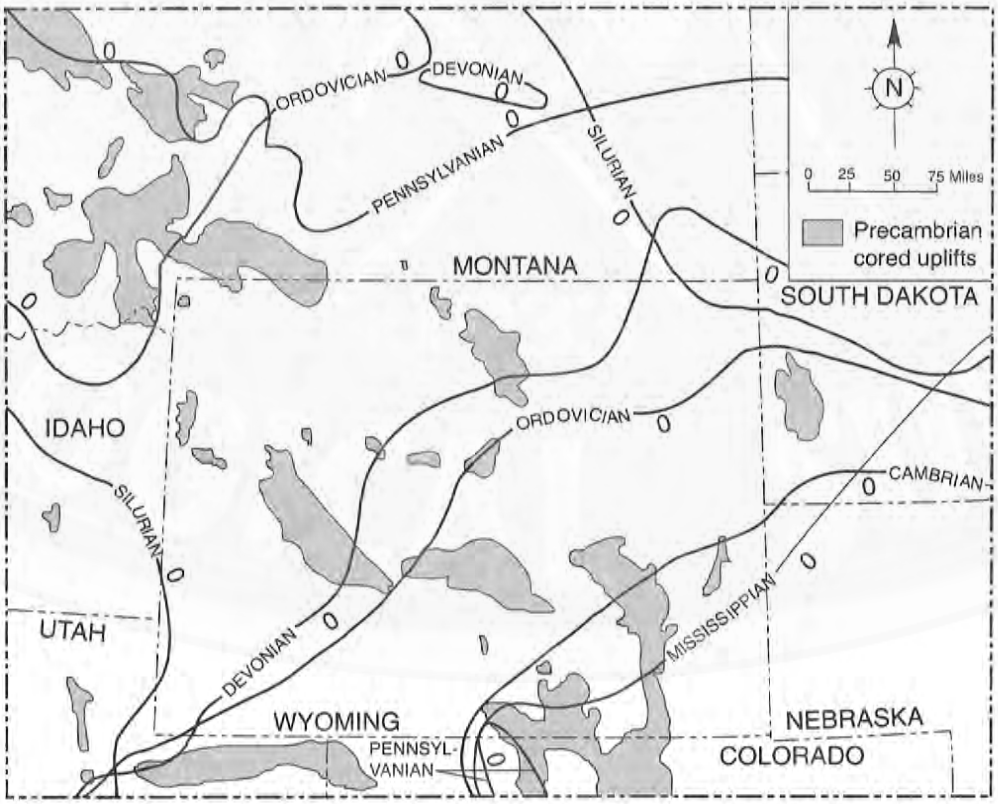


Figure 1. Distribution of Paleozoic zero isopachs in the Wyoming region. Permian strata were present throughout Wyoming at the close of the Paleozoic Era. (Modified from D.L. Blackstone, Jr. (1963). A zero has been added on the side of each line where rocks of that age are absent.) Incorporation of post-1963 information would change details, but not the basic pattern.

the Paleozoic history of southeastern Wyoming was provided by Sando and Sandberg (1987). The Paleozoic nomenclature and correlations shown in the 15 Wyoming columns of the American Association of Petroleum Geologists correlation chart for the central and southern Rockies region (Kent and others, 1988) differ only slightly from the terminology and age assignments utilized by Love and Christiansen (1985).

The annual guidebooks published by the Wyoming Geological Association since 1947 are a major source of information for those interested in the detailed geology of each Wyoming basin. For the regional picture outside Wyoming, the Rocky Mountain Association of Geologists' atlas cited above (Mallory, 1972) and volumes edited by Fouch and Megathan (1980) and Stewart and others (1980) are especially useful.

Cambrian Period

If we consider the Paleozoic record to have begun when the sea floor was first littered with tiny shells, then the place we call Wyoming was entirely above sea level at that time–a mosaic of ancient igneous and metamorphic rocks destined to remain in the erosional regime for another 30 million years. The terrain would have appeared desolate and inhospitable to a human eye, but there is no reason to believe it was lifeless. A microbial ecosystem involving photosynthetic and anaerobic bacteria no doubt flourished in the upper few millimeters of sediment, where the surface was not bare rock. Members of this microbiota functioned as binders of surface sediment as well as producers of various gases before the advent of more complex terrestrial communities.

Given its early Paleozoic location on the craton but near a passive continental margin, Wyoming differed greatly from the region now west of it in relative subsidence and in thickness of accumulating sediments. The Cambrian record demonstrates the hinge-like behavior of the western margin of the state, which separated the Wyoming platform from the developing miogeocline of Utah and Idaho. There, a thick record of Lower Cambrian marine strata accumulated while the transgressing sea moved toward Wyoming. Although the basal Cambrian strata lack age-diagnostic fossils, circumstantial evidence indicates that the shoreline had not reached Wyoming before Middle Cambrian time.

For the next 30 million years, the remainder of the Cambrian Period, the sea advanced sporadically eastward while the shoreline maintained a general north-south trend. These directional terms refer, of course, to present geography. This frame of reference is attractive because of its familiarity and because the Paleozoic geography is so imperfectly known. It must be remembered, however, that interpretations of Cambrian geography typically picture Wyoming near the equator and rotated 60° or more dextrally from its present orientation (Scotese and McKerrow, 1990).

By the end of the period, the Precambrian rocks were blanketed by siliciclastic deposits (**Figure 2**) whose progressive decrease in age from west to east reflects their transgressive origin. The surface across which they spread was dimpled with low hills–a few stood as much as 300 feet above the deepest valleys–so at any one time the advancing strandline formed an irregular border to a seascape dotted with islands.

The blanket of pebbles, granules, and sand reached a maximum thickness of between 500 and 600 feet. Lateral variation in thickness of the Flathead Sandstone, as these deposits are termed in most of the



Figure 2. Nonconformity between Middle Cambrian crossbedded sandstone (Flathead Sandstone) and Precambrian granite, Wind River Canyon, north-central Wyoming. Hammerhead for scale on contact at left center.

state, reflect both the irregular surface on which it accumulated and its intertonguing relationship with the dominantly shale sequence above it. Typical Flathead outcrops are reddish brown conglomerates and sandstone with abundant cross stratification in various styles. Most of these rocks are compositionally and texturally mature, but arkosic beds are present. Skeletal material is rare, although trilobite parts and inarticulate brachiopods have been found. Vertical dwelling burrows are abundant in some units.

A detailed analysis of bedding and lithology of Flathead strata (Middleton and others, 1980) indicates that most of the formation originated in nearshore and shallow shelf depositional environments. Common bedforms imply origin as subtidal sand-wave complexes migrating offshore under the influence of storm-surge and tidal currents. Other forms and orientations of cross stratification represent local preservation of beach and intertidal sandflat deposits. By contrast, the sedimentological immaturity and bedding geometry of basal strata in south-central Wyoming suggest an origin as transverse bars and channel deposits in braided streams.

Precambrian rocks in northeastern Wyoming were finally covered by sand in Late Cambrian time. These deposits are referred to the Deadwood Formation, a name that originated during early work on the stratigraphic sequence flanking the Black Hills. By the time the Deadwood siliciclastic deposits were accumulating, the Flathead Sandstone of western Wyoming was buried beneath hundreds of feet of shale and limestone (**Figure 3**). Just as grain size typically decreases from bottom to top in Flathead strata at any one locality, so the overall Cambrian record in western Wyoming is a fining-upward sequence. Two hundred or more feet of coarse clastic rocks of the Flathead are overlain by an 800-foot section dominated by shale (Gros Ventre Formation). The succeeding 300-foot interval, representing the rest of the Cambrian record in this area, consists mainly of limestone (Gallatin Limestone).

The simple picture of a transgressive sequence, in which superposed lithologies represent deposition successively farther from shore, is valid as a generalization, but stratigraphic details dictate a more complicated story. Limestones interrupting the Gros Ventre shales imply recurrent changes in pace of the transgression; shales in the Gallatin Limestone interval and minor disconformities record temporary retreats of the sea.

Although western Wyoming was farther and farther seaward during successive transgressive maxima, the Cambrian record there is not one of progressively deeper marine environments. At times when the advancing shoreline placed this area beyond reach of land-derived sediment, carbonate production in the clear, warm water could more than compensate for the loss of siliciclastic material. Such times were increasingly common during the Late Cambrian, and carbonate shoals and intertidal flats veneered with algal mats occasionally developed seaward of the shallow subtidal seafloor undergoing mixed carbonate-siliciclastic sedimentation.

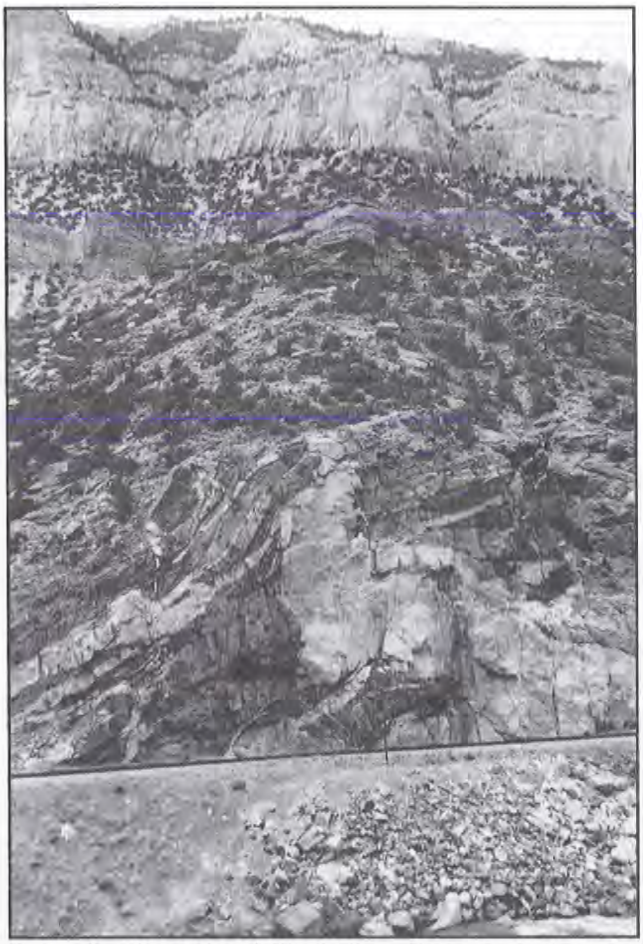


Figure 3. West side of Wind River Canyon, north-central Wyoming. Precambrian metamorphic rocks and granitic dike (railroad cut) overlain by Cambrian sandstone, shale, and limestone (tree-covered slope); white cliffs at top are Ordovician Bighorn Dolomite overlain by Mississippian Madison Limestone.

The record of the carbonate buildups is in the limestone ledges and cliffs that dominate Upper Cambrian outcrops in western Wyoming. These rocks are composed mainly of micrite and disarticulated parts of trilobites, echinoderms, and brachiopods. Grains of glauconite are common, and form the bulk of some beds. Prominent stromatolites cap a limestone member of the Gros Ventre Formation and these distinctive structures reappear in the Gallatin limestones. In both

formations, beds of flat limestone pebbles are common and conspicuous.

The intraformational limestone conglomerates deserve special mention as perhaps the most distinctive rock in the Cambrian sequence. The tabular clasts of laminated micrite are typically an inch or less thick and several inches in maximum length. They form discontinuous beds, commonly less than a foot thick, interbedded with bioclastic limestone, calcareous siltstone, and green shale. Two generations of clasts, distinguishable by degree of rounding and oxidation, are present in some layers. Clast orientation varies from random to strongly imbricate–the edgewise conglomerates of some authors. Interpretations of origin typically invoke early hardening of thin layers of lime mud by desiccation on tidal flats or by subaqueous cementation. In either case, the brittle layers were susceptible to fragmentation by storm-generated currents that tumbled the pieces and finally concentrated them in a layer as hydraulic energy waned.

Stratigraphic nomenclature for Wyoming's Cambrian strata consists mainly of names that originated in southwestern Montana (e.g., Gallatin Group) and northwestern Wyoming (e.g., Gros Ventre Formation). In these areas, the basal coarse-clastic section (Flathead Sandstone) is overlain by a succession of thick units, each of which represents a period when either clay and silt deposition or carbonate sedimentation prevailed. Over the years, names have been proposed for each major unit and for combinations thereof. Thus, the Gros Ventre Formation overlying the Flathead is a package of three members (shale-limestone-shale) followed stratigraphically by two predominantly carbonate formations comprising the Gallatin Group.

The body of literature dealing with these rocks presents a kaleidoscopic scene, in which familiar names repeatedly change in rank (member, formation, or group) or undergo expansion or contraction in stratigraphic and geographic limits. The nomenclature problem is exacerbated by the fact that the limestone intervals give way to shale and siltstone to the east and south. As the carbonate units lose their identity in these directions their names become inapplicable, and placement of formation and group boundaries that is easily done in northwestern Wyoming becomes very subjective. As might be expected, several names have been introduced for parts of the post-Flathead sequence of central Wyoming. Their lack of general acceptance is symbolized by their absence from the explanation sheet accompanying the latest geologic map of Wyoming (Love and Christiansen, 1985).

Cambrian biostratigraphy is based on a succession of trilobite assemblages, of which certain genera have proven to be more valuable than others. Although they are in no sense evenly distributed in Wyoming Cambrian strata, trilobites have provided the basis for relating the mosaic of lithostratigraphic units to the series and stages of the standard Cambrian time scale. For example, the first appearance of certain taxa in the upper Gros Ventre Formation dictates placement of the boundary between Middle and Upper Cambrian within the formation rather than at its base or top.

A search for trilobites produces only frustration at many outcrops, whereas their skeletal debris is a major component of the rocks elsewhere. Complete skeletons are rare; the components typically came apart during molting or, to a greater extent, through postmortem action of scavengers and currents. Fortunately, many genera can be recognized by the morphology of one part of the original skeleton, typically the cranidium.

The succession of trilobite zones recognized in the Cambrian of the western U.S. has been established empirically, but the reason for recurrent extinctions is not understood. A few of these biostratigraphic boundaries are especially dramatic, in that a preexisting assemblage is replaced by a new one within a stratigraphic interval of inches. A detailed study of one of these boundaries in central Wyoming (Chronic, 1988) indicates that the extinctions and first appearances of trilobite taxa coincided with a modest increase in overall faunal abundance and diversity. This suggests that a subtle change in water temperature or chemistry was fatal to the preexisting trilobite assemblage, while favoring other species that were adapted to the new conditions.

By the end of the period, Cambrian strata formed a continuous cover over Wyoming with the possible exception of the southeastern area that borders the Transcontinental arch. This term is commonly used for a broad belt, characterized by positive crustal behavior, extending between Minnesota and the New Mexico-Arizona area. Although its relief was always subdued and its borders ill defined, the feature influenced lithofacies, biofacies, and thicknesses of stratigraphic units to varying degrees during much of the Paleozoic Era.

Maps showing lithofacies and paleogeography of the Wyoming region at various times in the Cambrian are featured in several publications by Lochman-Balk (e.g., 1972). Middleton and others (1980) provided an excellent review of Wyoming Cambrian sedimentology, with emphasis on the siliciclastic rocks. An example of detailed interpretation of Cambrian carbonates is that of Martin and others (1980), whereas Chronic's (1988) documentation of faunal changes bracketing one extinction event illustrates the type of detailed work needed in Cambrian biostratigraphy.

Ordovician Period

A hypothetical geologist whose knowledge of lower Paleozoic stratigraphy was limited to the Wyoming record might be excused for questioning the applicability of Sloss' (1963) sequence scheme. That seminal paper, based on craton-wide analysis, emphasized the separate identities of two major packages of lower Paleozoic strata termed the Sauk and Tippecanoe sequences. The former includes Cambrian and Lower Ordovician beds, whereas the latter consists of Middle Ordovician through Lower Devonian strata. The two are bounded, as are the other sequences in Sloss' stratigraphic synthesis, by craton-wide unconformities.

The only places in Wyoming where one can find a record of continuous deposition across the Cambrian-Ordovician boundary are in the northeastern and north-central parts of the state. There, a minimal record of Lower Ordovician strata is present in the upper parts of the Deadwood and Gallatin formations. The sandstones and limestones that yield Lower Ordovician conodonts are lithologically like those comprising the lower parts of the two formations; the systemic boundary is intraformational and conformable. Elsewhere in Wyoming where the two systems are in contact, a disconformity separates Upper Ordovician strata (typically) or Middle Ordovician strata (less commonly) from the underlying Cambrian.

Although the Lower Ordovician evidence in the upper Deadwood and Gallatin formations is inconspicuous and geographically localized, it contributes to the regional paleogeographic picture dictated by evidence from surrounding states. The combined evidence supports the concept of continuous marine sedimentation from Late Cambrian through Early Ordovician time over much of Wyoming.

The state preserves little evidence of the sand that spread southward over much of the central and western U.S. (e.g., St. Peter Sandstone) in Middle Ordovician time. A 25-foot thickness of red and white sandstone unconformably overlying the Gallatin Formation in northern Wyoming is correlated with the Winnipeg Sandstone of the Williston Basin and the Harding Sandstone of Colorado. It resembles the latter unit in yielding abundant small fragments of dermal plates of primitive fish. There is no way to know how much of this sandstone was eroded before deposition of the Bighorn Dolomite, which unconformably overlies it and extends far beyond its southern limit.

The new depositional phase, represented by the Bighorn Dolomite, was associated with the Late Ordovician transgression generally regarded as the greatest inundation in North American history. With the sand supply effectively stopped, the tropical sea covering Wyoming was floored by carbonate sediments. The mud was littered with skeletal material produced by diverse communities of brachiopods, rugose and tabulate corals, large orthocone nautiloids, gastropods as much as half a foot in diameter, and the picturesque receptaculitids currently interpreted as skeleton-building algae.

The Bighorn Dolomite forms nearly all of the 500 feet of Ordovician strata preserved in northern Wyoming. The formation is fairly uniform in degree of dolomitization, but four members have been differentiated on other characteristics. Toward the bottom of the formation, the dolomite commonly grades downward into a few feet of sandstone. This basal member, known as the Lander Sandstone, contains fossils that place it with the rest of the Bighorn as Upper Ordovician. It can, however, be confused with the Winnipeg, from which it was probably derived. The two sandstones are separated by a disconformity, but the Lander is present in some parts of northern and western Wyoming beyond the limit of Winnipeg preservation.

The prominent cliffs formed by the Bighorn Dolomite (**Figure 3**) in northern and western Wyoming reflect the massive character of the thick (250 feet) carbonate member overlying the Lander. Weathered faces (**Figure 4**) of this distinctive unit, the lowest of three carbonate members, exhibit a complex network of ridges. Of various theories proposed to explain the distinctive relief, the most likely one invokes different resistance to weathering on the part of abundant burrow fillings and intervening matrix.

Significant intercrystalline porosity makes the formation a reservoir in several oil fields of the thrust belt and the Bighorn Basin. The present areal extent in Wyoming of the several carbonate members of the Bighorn Dolomite is the result of post-Ordovician erosion, which beveled the thick carbonate sequence from north to south. Ordovician rocks were completely removed from southeastern Wyoming except for carbonate blocks that foundered in small kimberlite diatremes near the Colorado border.

The fossils reported by Chronic and others (1969) from the diatreme blocks include a diverse, normal-marine, Late Ordovician fauna dominated by brachiopods and corals. The faunal lists are less conclusive in supporting the theory that older Ordovician sediments also covered the area. Diatreme fossils with affinities to known Early Ordovician taxa include hyoliths, a small nautiloid, a gastropod, and a trilobite segment. A piece

of sandstone yielded a fragment of a fish plate like those found in Middle Ordovician rocks.

The reader will find more detailed regional reviews of Ordovician history in publications by Foster (1972) and Ross (1976).



Figure 4. Part of Bighorn Dolomite (Ordovician) exposed in northwestern Bighorn Mountains. Jacob staff in left foreground rests against typical fretted weathered surface.

Silurian Period

On a pleasant summer day in 1954, Clinton (Nick) Ferris, Jr. stepped out of a southern Wyoming hay meadow and into the lore of Rocky Mountain paleontology. Taking a break from the haying operation on his family's ranch, young Ferris, then a geology student at Colorado College, indulged his interest in rocks by collecting fossils from a boulder-strewn slope near the meadow (**Figure 5**). The site subsequently gained renown among regional paleontologists and stratigraphers as the Ferris outlier, the only Silurian rocks for 300 miles in any direction.

The circle of professional interest broadened dramatically when evidence of ultramafic rock caused the fault-block explanation for the site to be replaced by the kimberlite diatreme theory now accepted without question. With diamonds as an added incentive for prospecting, over 100 diatremes have been found in the Laramie Mountains of southeastern Wyoming and the Front Range of northeastern Colorado (Hausel, 1985). Very few of them have sedimentary inclusions, so one of the great ironies in the story is that discovery of the only diamond-bearing localities in the western U.S. was triggered by a fossil collection.

Before Ferris' discovery and the subsequent fossil identifications by John Chronic and other paleontologists, paleogeographic maps for the Silurian showed Wyoming and Colorado as part of an extensive area not covered by the sea (**Figures 1** and **6**). One could not ask for a better example of how a dramatic change



Figure 5. Ferris diatreme and surroundings, south of Laramie. Diatreme rocks form the sagebrush-covered knoll immediately beyond the pond, extending from the hay meadow (left) to the right edge of the picture. Hills in the upper half of photo and the steep slope between camera and pond are Precambrian granite.

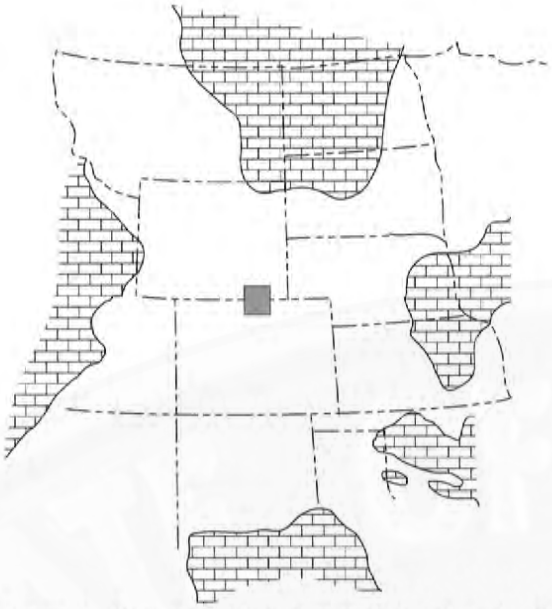


Figure 6. Regional distribution of Silurian rocks as known before diatreme discovery (brick pattern). The small rectangle on the Wyoming-Colorado border shows the location of the first diatremes discovered. (Reproduced from Chronic and others, 1969.)

in conventional paleographic wisdom can be required by a discovery at a single locality.

The Silurian brachiopods and corals reported by Chronic and others (1969) from the Ferris diatreme and a smaller one nearby represent at least two ages (late Llandoverian and late Wenlockian or Ludlovian). This fact, together with the environmental significance of the fossils and their carbonate host rock, leaves no doubt that this area was beneath clear-water seas for a significant part of Silurian time. It follows that the same was probably true for much or all of Wyoming.

In map view, the Ferris diatreme occupies an oval area with maximum length of about 250 feet. Like all those subsequently discovered in the Front Range, it is surrounded by Precambrian crystalline rocks. The closest outcrops of in situ strata are about 2 miles away, where Pennsylvanian arkoses cover the granite. In contrast to the absence of sedimentary rock at the vast majority of the diatremes, the Ferris outcrops consist of the foundered blocks of Ordovician and Silurian carbonates (**Figure 7**). Cuttings from the bottom half of a 230-foot borehole drilled at its center indicate that ultramafic matrix dominates over inclusions at that depth (Chronic and others, 1969).

It is not known whether the diatreme intrusions reached the Earth's surface, nor is it certain when they occurred during the window of opportunity bracketed by origin of the youngest fossiliferous blocks and deposition of the Pennsylvanian arkose. Absence of undoubted Devonian and Mississippian inclusions favors late Silurian or early Devonian as the most likely time of intrusion. Other than the diatreme material, the only certain Silurian rocks in Wyoming are along the western border in the form of the easternmost record of the Laketown Dolomite.

Devonian Period

Wyoming's position on the craton margin was not favorable for a detailed recording of the numerous rises and falls of Devonian sea level now being documented by stratigraphic studies farther west. Only the



*Figure 7. Largest outcrop of sedimentary rock in the Ferris diatreme (visible at right center of **Figure 5**). Hay meadow (upper left) and hay crib (upper right) are also shown in **Figure 5**. This outcrop contains Ordovician fossils, whereas nearby boulders yield Silurian brachiopods. (Hat and camera bag for scale.)*

major transgressive pulses affected the state, and long regressive intervals resulted in loss of strata deposited not long before. Those units that were preserved typically change thickness laterally because of both depositional thinning in the direction of transgression (west to east, and north to south) and differential stripping during times of subaerial erosion.

The period began with Wyoming emergent and vegetated with primitive vascular plants. Silurian strata were lost to erosion, and drainage systems developed on older bedrock. As far as can be told from the rock record, the maximum Early Devonian transgression onto the craton affected Wyoming only by turning the seaward ends of the valleys into estuaries. Sediments that accumulated in several of these are preserved as isolated lenses, as much as 75 feet thick, along the disconformable top of the Bighorn Dolomite in northern Wyoming.

Collectively known as the Beartooth Butte Formation, the lenses of conglomerate and carbonaceous dolomite are the only Lower Devonian rocks identified in Wyoming. The bizarre fossil assemblage obtained from them, including eurypterids, vascular plants, and fish, signifies their estuarine origin.

Wyoming was exposed to erosion during a major Middle Devonian regression that affected many parts of the world. When a Late Devonian sea encroached on the state, the initial depositional environments replicated the estuarine conditions of the Beartooth Butte Formation. The Upper Devonian channel deposits can be confused with their Lower Devonian counterparts in northern Wyoming, where both can be found along the disconformity at the base of the Devonian sequence.

The advance of Late Devonian sea water into the valleys draining the terrain of Bighorn Dolomite bedrock was the initial expression in Wyoming of a major rise in sea level. At the height of this transgression, marine conditions prevailed throughout the northern and western parts of the state. The resulting deposits consist of several hundred feet of sparsely fossiliferous carbonate rock, commonly dolomitized, with interbedded siltstone and sandstone. Montana nomenclature (Jefferson Limestone and Three Forks Formation) is applied to these Upper Devonian beds in northern Wyoming, whereas equivalent strata in the western part of the state are commonly called the Darby Formation. In the latter unit, beds of siltstone and sandstone become increasingly common eastward relative to dolomite intervals; the shore may not have been far beyond the present erosional limit of the formation.

Vacillation in water depth and supply of noncarbonate sediment during deposition of the Jefferson-Three Forks interval resulted in differences in bedding and lithology that led geologists in southwestern Montana to distinguish two members in each formation. Although carbonate rocks are dominant, mudstone is common in the upper part of the interval and an evaporitic facies a few tens of feet thick is present along the Idaho-Wyoming border.

The unconformity at the top of the sequence bevels it from northwest to southeast in such a way that successively older members cap the sequence southward from the state line to its southeastern limit. There, the remaining beds are inconspicuous and easily overlooked or misidentified.

Although the regional disconformity at the top of the Jefferson-Three Forks and Darby beds seems to signify the upper limit of the Devonian record in the state, this is not literally true. Late Devonian conodonts are present in a few tens of feet of dark gray shale and silty dolomite above the disconformity (Sandberg and Klapper, 1967). These unusual strata, constituting the basal member of the Madison Limestone, encompass the systemic boundary; Early Mississippian conodonts have been obtained from the upper beds.

To the southeast, a similar conodont succession places the Devonian-Mississippian boundary in a thin unit not far above Precambrian granite. In this area, the north end of the Laramie Mountains and vicinity, the conodonts are in siltstone and sandy dolomite separated from the crystalline basement by a variable thickness (6 to 186 feet) of resistant sandstone. In southeastern Wyoming, any such outcrop of unfossiliferous sandstone nonconformable on Precambrian granite is fair game for speculation as to geologic age. Nominations can start with the Cambrian, but evidence in this case favors Devonian.

The above evidence indicates that marine conditions had returned to Wyoming before the Devonian Period ended, but it does not imply that sea water covered the state at that point. Lithologic differences and lack of stratigraphic continuity between pertinent strata in northwestern and southeastern Wyoming suggest that the latest Devonian marine onlap did not completely cover central and southern Wyoming.

A review of the Devonian System in the Rocky Mountain region by Baars (1972) includes several thickness and lithofacies maps. Johnson and others (1988) offered a detailed interpretation of Early Devonian history of the western U.S., and Sandberg and others (1988) did the same for the Late Devonian. All three of these sources cite many references that deal specifically with Wyoming.

Mississippian Period

Far west of Wyoming, the plate margin changed from the passive behavior of its early Paleozoic history, and an allochthon of pre-Mississippian oceanic strata moved onto the continent margin. Although sediments derived from the resulting Antler orogenic belt did not reach the Wyoming shelf, the tectonic activity to the west may have influenced the advances and retreats of the sea recorded in the thick Mississippian carbonate sequence of Wyoming.

The Mississippian Period was characterized by carbonate deposition over a major part of North America. Wyoming's heritage from these warm, clear-water seas was a wedge of carbonate strata (**Figure 8**) thickening irregularly west and north from a zero edge in the southeast to more than 1000 feet at the western border. Intercrystalline and vuggy porosity in some of these rocks makes them important as major aquifers and as oil and gas reservoirs in areas such as the Bighorn Basin.

Over most of the state, the carbonate strata are given formation rank as the Madison Limestone–a somewhat misleading term considering the extent of dolomitization. All but the uppermost of six formally designated members in central Wyoming are carbonate. They differ in ratio of dolomite to limestone, bedding type, texture, grain origin, and chert content. The Madison is treated as a group in western Wyoming, where two Montana names–Lodgepole Limestone and Mission Canyon Limestone–are applied to the lower and upper parts, respectively, of the thick section. At the other side of the state, the Black Hills term, Pahasapa Limestone, is employed for Madison-equivalent strata flanking the Black Hills, and correlative strata in the Hartville uplift to the south are called the Guernsey Formation.

As implied in the comments on Late Devonian history, the long interval of marine conditions represented by Madison strata began rather tentatively. The narrow seaway that extended into part of Wyoming in latest Devonian time was reestablished very early in the Mississippian Period after a brief absence. The limited areas drowned in these incursions received



Figure 8. Wind River Canyon (looking south) in north-central Wyoming. Cambrian rocks form slopes rising from the river; Bighorn Dolomite (Ordovician) and Madison Limestone (Mississippian) form cliffs. Silurian and nearly all Devonian strata were removed by pre-Madison erosion. The Wind River is bordered by a highway (left) and a railroad (right).

several tens of feet of conodont-bearing dark shale and silty dolomite now recognized as a basal member of the Madison sequence. A disconformity separates Devonian and Mississippian parts of the member where both are present. In some areas, only the upper tongue is found, and the member is missing entirely in parts of western and north-central Wyoming.

The shallow-bay environment terminated abruptly as a major eustatic rise produced a rapid transgression over the craton, which left only an area along Wyoming's southern border emergent. To the west, the basal black shale member of the Lodgepole Limestone grades upward into silty, variably cherty limestones of its middle member. At the western border of the state, water depths during deposition of these beds may have reached several hundred feet at times. Oxygen-deficient conditions below the photic zone are suggested by laminated rock with no algal remains and only rare shells. Correlative shelf strata in central Wyoming are now extensively dolomitized.

The thin beds, typically rich in skeletal debris, that typify the upper half of the Lodgepole Limestone represent a time when carbonate accumulation was catching up with sea-level rise. Repetition of small-scale, shoaling-upward cycles (mudstone-packstone-grainstone) reflect migration of tidal bars of oolitic and skeletal sand across the mud bottom. Dense populations of bryozoans and crinoids thrived in deeper water between the shoals, and grains in these less disturbed areas were typically micritized by endolithic algae. High-energy conditions were common in the shallower water to the east, as attested by cross-bedding, oolite, and abundant skeletal debris. Such indicators are common in central Wyoming in the part of the Madison Limestone equivalent to the much thicker upper Lodgepole.

A second major sea-level rise was accompanied by progradation of the shelf until the shelf edge coincided with the present western border of Wyoming. Aggrading calcarenites and lime mud in this area formed the Mission Canyon Formation. The interbedding of these facies reflects rapid alternation of turbulent and less agitated environments as fluctuation in depth of wave action or shifts in tidal currents winnowed the mud sea floor. The reworked skeletal material was concentrated in shifting tidal bars, where smaller grains became nuclei for oolitic coatings.

Development of extensive shoals at the platform edge impeded water movement onto the shelf. Lagoonal conditions developed there, and the restricted circulation culminated more than once in precipitation of evaporites. When dissolution removed these beds from the sequence much later, collapse of superjacent rock produced layers of breccia. Two such zones, each several feet thick, are found in the Mission Canyon Limestone. Similar evidence of evaporite dissolution is present in the laterally equivalent part of the Madison Limestone including the two upper members of the formation in central Wyoming–cherty dolomite overlain by red shale and stromatolitic carbonate.

The fossil content of Madison rocks is of interest in several respects. Some beds are virtual coquinas, whereas others are barren of fossils or exhibit only scattered crinoid columnals. The overall macrofauna includes solitary and colonial corals, fenestrate and nonfenestrate bryozoans, brachiopods of several orders, gastropods, trilobites, and plates and spines of various echinoderms. Conodonts and foraminifera are present in many beds, and thin sections of bioclastic rock commonly exhibit skeletal fragments attributable to green and red algae. Correlation of various parts of the Madison sequence with series and stages defined in the Mississippi Valley and Europe is based on conodonts and endothyroid foraminifera. Although more provincial, corals provide a zonation useful in correlating Madison strata with Mississippian units elsewhere in western North America.

Both rugose and tabulate corals are common in Madison strata but no reefs are present. These corals lived as solitary polyps and scattered colonies on loose sediment. Unlike many of their modern relatives, their growth habits did not involve lateral expansion by encrustation, and they did not form hard substrates above wave base. Also lacking in Wyoming Mississippian strata are Waulsortian mounds of the classic type described from outcrops of Mississippian and Lower Carboniferous outcrops elsewhere. Development of these enigmatic lenticular buildups of fossiliferous lime mud seems to have been favored by slope environments below greater water depths than those represented by most Madison strata.

Addition of strata to the Wyoming shelf was abruptly halted by a major regression that moved the shoreline to the western border of the state, exposing the entire platform to subaerial processes. Abundant meteoric water leached the surface, forming caverns, sinkholes, and breccia (**Figure 9**), while subsurface evaporite layers were dissolved by ground water. West-flowing rivers draining the old highland along the southern border of the state built a delta southwest of Wyoming. Longshore currents brought arkosic sand north from that area and it was interbedded with carbonates that continued to accumulate in shallow water just offshore from the karst plain (DeJarnett, 1984).



Figure 9. Paleokarst on white Mississippian limestone (Guernsey Formation) overlain by red Pennsylvanian sandstone (Hartville Formation). Cliff overlooks the North Platte River near Guernsey in eastern Wyoming.

Wyoming's well-developed karst topography was probably mantled by lateritic soil supporting a low-latitude plant community that provided habitat for insects and primitive amphibians. The picture seems reasonable, but there is no direct evidence other than red mudstone in some of the sinkholes; the rest was removed as the scene changed again to a seascape.

The final advance of the Mississippian sea affected only the western half of the state, where it deposited the first of three diverse members known today as the Amsden Formation. The foundation was thus laid for one of the most vexing controversies in Wyoming stratigraphy—the ages and lateral relationships of the various parts of that formation.

The basal member, the Darwin Sandstone, represents a dramatic change from the carbonate environments that produced the underlying Madison Limestone. The white, crossbedded sandstone is more than 150 feet thick in places. Its textural and compositional maturity implies that it is recycled material, possibly delivered by rivers draining lower Paleozoic sandstones far to the northeast. The unstable substrate in the advancing sea that reworked the sand was inhospitable for organisms, and the unfossiliferous aspect of the Darwin necessitates an age assignment based on circumstantial evidence. Although there is general agreement that the Darwin was deposited late in the Mississippian Period, there is no such consensus as to whether or not deposition was continuous from Mississippian into Pennsylvanian time.

The huge body of information compiled by Craig, Connor, and others (1979) on the Mississippian System in the U.S. puts the Wyoming record in continent-wide perspective. Maughan (1979) and Sando (1979) offered frank appraisals of their different interpretations of Amsden and related strata. A chronology of Devonian and Mississippian eustatic changes and epeirogenic events affecting the western U.S. as far east as Wyoming was compiled by Sandberg and others (1982) in a paper that includes five Mississippian paleogeographic maps. DeJarnett's (1984) detailed study of a thick Mississippian section in the thrust belt documented various biofacies and lithofacies that are widespread in rocks of the same age in western Wyoming.

Pennsylvanian Period

Had David Hume been a stratigrapher, he might have meditated that the existence of disconformities, as well as beauty, is in the mind of the observer. He might, that is, if he were a biostratigrapher arguing the "Amsden problem" with a physical stratigrapher. Although both camps cite faunal and physical evidence to support their contradictory views, the conclusion reached seems to depend on the weight given to certain fossils, as indicators of Mississippian age, versus lithostratigraphic criteria, as evidence for regional disconformities.

The current controversy centers on the relationship between the Amsden's basal member (Darwin Sandstone) and the two overlying members (Horseshoe Shale and Ranchester Limestone). One view holds that the Darwin is separated from the others by a regional disconformity and that Horseshoe strata are the initial deposits of a later transgression that initiated Pennsylvanian sedimentation in the area. The alternative view interprets the Horseshoe and Ranchester fauna (foraminifera, ostracodes, corals, brachiopods, and mollusks) to include Late Mississippian forms in the west and Early Pennsylvanian ones in central Wyoming. In this view, the Darwin-Horseshoe contact is conformable, and the three members are parts of a single transgressive episode that did not reach eastern Wyoming until Pennsylvanian time.

Part of the controversy concerns the relation of Amsden-like Mississippian strata in westernmost Wyoming to the Amsden of central Wyoming. If strata of the two areas were originally continuous, the transgressive theory is correct. The opposing school, however, maintains that the Mississippian wedge is overlain unconformably by Pennsylvanian strata. If so, similarities between units in that wedge and rocks of the Horseshoe and Ranchester members are coincidental and not grounds for postulating a time-transgressive relationship. The Amsden controversy is by no means the only one involving Wyoming Paleozoic rocks, but there is probably no better example of

the problems confronting stratigraphers who deal with poorly exposed, sporadically fossiliferous, and heterogeneous formations.

The shelf conditions that characterized Wyoming during the Pennsylvanian contrasted sharply with localized areas of rapid subsidence to the west and south (e.g., Sublett Basin of Idaho, Oquirrh Basin of Utah, central Colorado trough) and the rising mountains (e.g. ancestral Front Range uplift) in Colorado. The shelf was not, however, tectonically inactive. The extraordinary crustal instability to the west and south was mimicked in miniature by subtle warping of the platform as the period progressed. The positive areas that developed from time to time affected both sedimentation and preservation of the record. These aspects of the state's late Paleozoic history are discussed by Maughan (1993) in his paper on the Ancestral Rocky Mountains.

The dominant role of terrigenous detrital sediment in Wyoming's Pennsylvanian strata, about 500 feet thick on the average, represents a major change from the Mississippian, when carbonate factories had little competition. In the Pennsylvanian, limestones formed from time to time but, with the exception of the Ranchester sea, clear-water conditions were typically brief. Their influence is most notable in the northeastern and eastern parts of the state, where the prominence of limestones in interbedded siliciclastic-carbonate units attests to the proximity of open seas in those directions.

Although opinions differ as to whether or not Wyoming was completely emergent at the beginning of Pennsylvanian time, all agree that shallow marine conditions were present early in the period. Mild crustal warping had produced a positive area along the Idaho border, and the northern end of the ancestral Front Range extended into south-central Wyoming. As a result, access of the Early Pennsylvanian sea to the Wyoming shelf seems to have been from the north, southwest, and southeast.

Today, the red mudstones of the Horseshoe Shale Member, commonly 100 feet thick, produce a distinctive zone of colorful soil separating ledge-forming Mississippian units below from the overlying Ranchester Limestone. The redbed depositional environments involved frequent flooding of coastal mudflats, but only rarely did an inundation persist long enough in central Wyoming to generate a thin limestone bed. Interbedded carbonate increases to the southwest, where the Horseshoe Shale grades into the Round Valley Limestone of Utah.

Expansion of the sea from the west into Wyoming resulted in the relatively clear-water conditions in which 100 feet or more of cherty carbonates with red mudstone partings were deposited (Ranchester Limestone Member). This interval of dominantly carbonate deposition was terminated by a major influx of siliciclastic sediments that comprise the bulk of the Tensleep Sandstone (**Figure 10**), one of the most important petroleum-producing formations in the state. The 200- to 300-foot thickness typical for the Tensleep includes numerous dolomite layers, some as much as 12 feet thick. Such units at the top of the formation can be confused with overlying Permian formations. Close-spaced carbonate layers at the base of the section pose a different problem, in that they are included in the Tensleep by some workers and in the Amsden by others.

As the formation name implies, the dolomite beds are a minor component of the Tensleep. The thick body of quartz sand they interrupt exhibits prominent crossbedding–some sets are more than 50 feet thick–as well as distinctive intervals of large-scale contorted bedding. A prolific source must have existed in Mon-

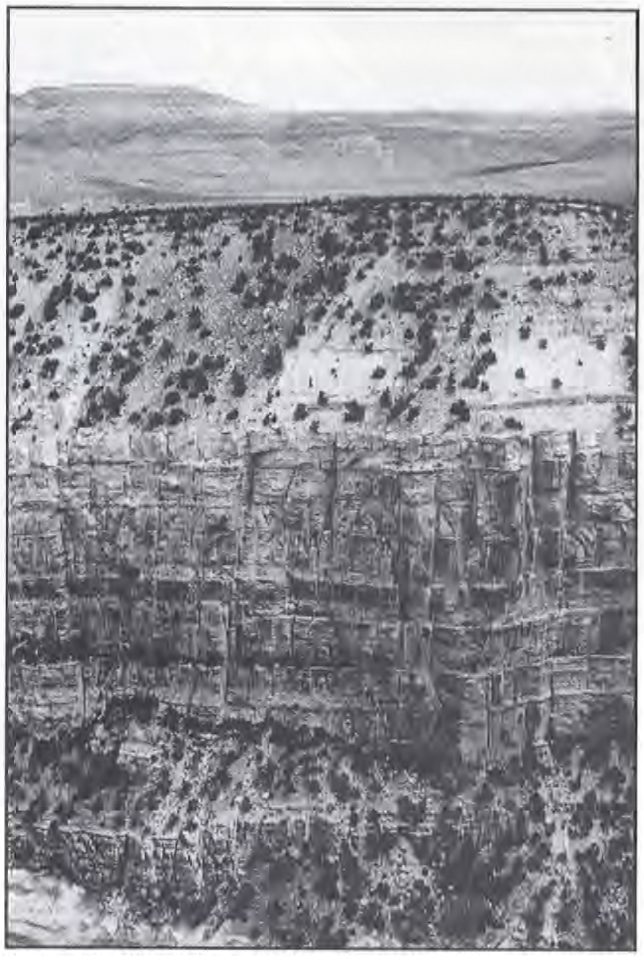


Figure 10. Cliff of Tensleep Sandstone (Pennsylvanian) at the north end of Wind River Canyon (east side). Tree-covered slope from cliff to canyon rim is formed by Park City Formation (Permian). (Highway at lower left corner.)

tana, northern Idaho, or the Canadian Shield to supply the great blanket of sand that spread over the Wyoming shelf, as well as adjacent areas where other formation names are applied (e.g., Quadrant in Montana, Wells in Idaho, and Weber in Utah).

Wind seems to have played a major role in the transport and deposition of the sand. Several eolian dune types, differentiated by distinctive bedding geometry and grain characteristics, have been identified in the formation. Other intervals lacking an eolian signature probably represent reworking of previous deposits by shallow seas that occasionally flooded the area. The more extensive incursions produced the intercalated carbonate units. Such evidence of fluctuating sea level is not surprising for a period of Earth history when glaciers were waxing and waning in the southern hemisphere.

Another factor that probably influenced distribution of marine and nonmarine facies within the Tensleep was variation in relief of the platform surface. Depositional topography in dune fields may have been significant, but slight differences in epeirogenic activity in different parts of the platform were probably more important. Positive areas would have been the sites of the most continuous eolian activity, while adjacent areas were more frequently submerged.

In eastern Wyoming, marine conditions were more continuous and clear-water environments more frequent than was true for the area of Tensleep deposition. The hundreds of feet of limestone and sandstone that accumulated in this area are now exposed in the Black Hills uplift, in the smaller Hartville uplift to the south, and along the flanks of the Laramie Mountains. By 1908, a local name had been applied to these rocks in each of the three areas. From north to south, the synonyms, Minnelusa, Hartville, and Casper mask the overall similarity in age and lithology of the rocks to which they apply.

Bryozoans, brachiopods, mollusks, and echinoderm plates are found in many beds of these formations, but fusulinids are of special interest because of their biostratigraphic importance. The succession of fusulinid species and genera present in each of the three formations establishes their mutual age equivalency. It also demonstrates that all four Pennsylvanian provincial series, as well as the lowest Permian one, are represented in the eastern Wyoming sequence.

During much of the Pennsylvanian Period, the northern part of the ancestral Front Range greatly influenced sedimentation in southeastern Wyoming. Gravel, sand, and clay derived from the exposed granite accumulated as an apron, hundreds of feet thick, of alluvial fans and braided-stream deposits flanking the highland. These arkosic fluviatile sediments comprise the Fountain Formation (**Figure 11**). The interplay of fluctuations in both sea level and supply of terrigenous sediment led to complex intertonguing of marine and nonmarine facies where the Casper Formation gives way to the Fountain. As the period progressed, this transition zone migrated toward the source area in response to the decreasing supply of terrigenous sediment. As the Pennsylvanian Period came to a close, strata continued to accumulate throughout much of eastern Wyoming while subaerial erosion was already erasing the latest part of the Pennsylvanian record elsewhere in the state.

The monumental study of the Pennsylvanian System in the U.S. produced by McKee, Crosby, and others (1975) includes two sections of text dealing with Wyoming and adjacent areas, and numerous lithofacies, isopach, and paleotectonic maps for the country as a whole. Another valuable synthesis for the Rocky Mountain region, by Mallory (1972), is illustrated with unusually colorful lithofacies maps. The report on the Tensleep Sandstone by Kerr and others (1986) illustrates the sedimentological complexities of that important formation.



Figure 11. Nonconformity separating Proterozoic granite from Pennsylvanian fluviatile sandstone (Fountain Formation) at the Wyoming-Colorado border south of Laramie.

Permian Period

In the Permian world, the part of the crust under discussion here was a tiny part of the great supercontinent Pangaea. Assembly of this greatest of all Phanerozoic landmasses had placed much of the margin of the North American craton far inland, but this was not true for the part of the margin bordering Wyoming. As had been true throughout its Paleozoic history, most of the state behaved tectonically as a stable platform relative to the area now to the west, where passive-margin subsidence during the first half of the era was followed by more complicated tectonism.

As the period began, much of Wyoming was losing the last-formed part of its Pennsylvanian record. By contrast, Early Permian fusulinids in the upper beds of the Weber and Wells sandstones show that deposition continued along the state's western border, and similar evidence is found in the upper part of the Casper, Hartville, and Minnelusa Formations of eastern Wyoming. There, the marine presence resulted in an interbedding of dolomite and sandstone interrupted, in the Minnelusa, by evaporite beds.

The Permian history of the Minnelusa is of special interest to the petroleum industry. The upper member, some 200 feet thick, consists of eolian sandstones alternating with marine carbonates and evaporites. The sandstone units have significant relief on their upper surfaces, and commercial quantities of oil have been trapped in the topographic highs. Relief on the top of the formation was created by an interlude of subaerial erosion that preceded deposition of overlying Permian redbeds. Lateral changes in thickness of lower sandstone units may reflect dune field topography modified by subsequent transgressions.

Farther south, a vestige of a coastal dune field that bordered the eastern sea is exposed at the Wyoming-Colorado border south of Laramie. The picturesque crossbedded sandstone, of special interest as the type locality for the term "festoon crosslamination" (Knight, 1929), overlies Pennsylvanian fluviatile strata of the Fountain Formation. The sandstone, however, is laterally equivalent to nearby strata of the upper Casper Formation that contain Early Permian fusulinids. Large-scale contorted bedding in the sandstone (**Figure 12**) is remarkably like that in several other formations (e.g., Tensleep) and is no better understood here than in the other occurrences.

The subaerial erosion that affected the central part of the state during the Pennsylvanian-Permian transition expanded to include the eastern area of earliest Permian deposition, thus producing the economically important disconformity at the top of the Minnelusa Formation. This surface, as well as the entire Wyoming platform, was then subjected to aggradation until, by the end of the period, several hundred feet of Permian strata had been deposited over the main part of the shelf. This sequence accumulated as several contrasting depositional environments moved back and forth over the state and adjacent areas. The resulting complex of intertonguing rock units now exposed around Laramide uplifts in Wyoming, Utah, Idaho, and Montana has challenged stratigraphers of several generations. They, in turn, have created names for members, formations, and groups in such overlapping profusion as to render the word "nightmare" inadequate.

The basic picture involves deposition of redbeds (Goose Egg Formation) over the eastern half of the state while a peculiar sequence of phosphatic and cherty strata (Phosphoria Formation) accumulated along the western border. The intervening west-central area was characterized by carbonate deposition (Park City Formation), while sand spread into northwestern Wyoming from the north (Shedhorn Sandstone). The four formations represent end-member concepts more easily dealt with in theory than in practice. Nowhere in the state does one of them occur without including tongues of one or more of the others. At localities where tongues of two formations are equally prominent, the choice of one formation name for the section can seem very arbitrary. This has encouraged many geologists, particularly those in the petroleum industry, to disregard the term Park City Formation and to apply the term Phosphoria Formation as far east as the Goose Egg redbeds.

Carbonate members of the Park City Formation are important reservoirs in oil fields of the Wind River and Bighorn Basins. Furthermore, the concentrations of phosphorite in the western facies have been commercially significant for several generations. Given the



Figure 12. Contorted eolian sandstone (Permian part of Casper Formation) at the Wyoming-Colorado border south of Laramie.

economic motivation, it is not surprising that the Phosphoria and Park City Formations have been subjected to an unusual amount of study. It is safe to say that the resulting literature is more voluminous than that for any other Paleozoic formation in Wyoming.

The Phosphoria Formation of the Overthrust Belt is the eastern expression of unique depositional environments in a basin whose north-south axis was west of Wyoming. The long interval of sand deposition that had characterized the area during Pennsylvanian and earliest Permian time was terminated when an expansion of the sea halted erosion on the Wyoming platform. Basin flank and adjacent shelf were then blanketed with impure lime mud, which became the cherty and sandy dolomite comprising the basal member of the Park City Formation. Eastward of the carbonate limit, fine siliciclastic sediment covered the erosion surface. Thoroughly oxygenated during or soon after deposition, these layers became the first of the thick succession of Goose Egg redbeds.

The sedimentary history of the Phosphoria depositional basin is highlighted by two intervals when conditions favored accumulation of phosphatic sediments. The resulting stratigraphic units, the older Meade Peak (below) and younger Retort Phosphatic Shale Members, consist of black to brown shale and phosphorite. Beds of the latter vary in siliciclastic content and in texture. Some layers are oolitic, others are pelletal, and some consist mainly of fragmented valves of orbiculoid inarticulate brachiopods.

A completely satisfying explanation for the origin of the phosphatic members has yet to appear. Most interpretations invoke the upwelling of nutrient-rich waters from the depths of the basin onto the bordering ramp or shelf, and they typically envision high production of phytoplankton accompanied by slow sediment accumulation on an oxygen-deficient bottom. All workers agree that each of the two phosphatic members is part of a separate transgressive-regressive cycle, but opinions have differed as to whether or not deposition at the eastern limit of each member coincided with a highstand of the sea.

The two phosphatic members thin eastward to their distal edges in west-central Wyoming, where they are identifiable as thin beds in the Park City Formation. In basin sections where they achieve maximum thickness, the members come between and intertongue with equally enigmatic members composed of bedded chert and dark siliceous shale. Two concentrations of chert in different parts of the Phosphoria section have been formally designated as members. Their stratigraphic positions and dimensions, unlike those of typical lithologic units, may reflect localization of diagenetic rather than depositional environments. In any case, the chert members intertongue to the east with carbonate members of the Park City Formation (**Figure 10**), and both lithologies are interrupted by sandstone beds. These tongues of the Shedhorn Sandstone are commonplace in northwestern Wyoming, reflecting their Montana source.

The interval of sandy dolomite forming the base of the Phosphoria-Park City sequence is the lowest and least fossiliferous of the three carbonate members comprising the Park City Formation. While they were accumulating, the sea floor over much of the west half of Wyoming was formed of lime mud and skeletal debris. Percentages of these two components varied with time, as did the sites of thickest accumulation. The middle member is thickest in the west, where carbonate banks aggraded along the shelf margin. The thickest parts of the upper member are farther east on the shelf. Accumulating evidence indicates that lateral variations in thickness and carbonate facies of these members reflect the syndepositional influence of subtle structural highs and lows on the shelf.

The carbonate grains are typically skeletal fragments from bryozoans, brachiopods, and echinoderms. Ooids, pellets, and plates of phylloid algae are common in a few places. Packstone and wackestone fabrics predominate over grainstone, implying generally low-energy conditions in the carbonate-floored shelf seas. The type area for the term "fenestral fabric" (Tebbutt and others, 1965) is along the west flank of the Bighorn Mountains, where the distinctive rock is tens of feet thick in hogbacks formed by the upper Park City member. The same peritidal facies forms the petroleum reservoir in the highly productive Cottonwood Creek Field not far to the south.

With the exception of a gypsum quarry operated by a Laramie cement plant, one searches in vain for Wyoming sites where the Goose Egg Formation produces something of economic interest. The redbeds play a critical role in some oil fields, where they seal underlying Minnelusa reservoirs or block lateral migration of petroleum at carbonate pinchouts. This function, however, has not served to motivate sedimentological study; little such work has been done on the Goose Egg Formation (**Figure 13**) relative to the effort lavished on the economically important correlative formations to the west.

Detailed subsurface correlations of the Goose Egg and Park City Formations are facilitated by the vast eastward extent of several thin tongues of the latter formation. Each represents a brief time, presumably a highstand of the sea, when carbonate deposition interrupted the aggradation of fine siliciclastic sediments.

Three carbonate tongues from the Park City have been named, as has each of the redbed intervals they separate. Add another pair at the top, and the Goose Egg Formation consists of eight members. The upper two are generally considered to be Triassic, but are included in the formation because the package forms a convenient unit for surface and subsurface mapping.



Figure 13. Slope-forming red siltstone and carbonate ledge in Goose Egg Formation (Permian) overlying Tensleep Sandstone (foreground); west flank of Sheep Mountain anticline, northeastern Bighorn Basin.

The lowest carbonate tongue (Minnekahta Limestone) is especially famous for its areal extent; it has been traced in the subsurface as far as the western parts of Nebraska and the Dakotas. The transgressions that produced the carbonate members did not expand the habitat for the brachiopod-bryozoan communities thriving in the western half of the state. Only a few mollusks and fish probed the eastern environments, which no doubt became increasingly hostile in that direction. Some of the carbonate is stromatolitic, but much is nondescript dolomite that gives way laterally to evaporites. In fact, evaporite beds in the Goose Egg sequence typically correlate with, and are extensions of, carbonate members.

The great lateral extent of the thin carbonate-evaporite tongues implies that each originated in a standing body of water that had spread over a featureless surface essentially at sea level. The red clay-rich silt that forms most of the several hundred feet of Goose Egg strata apparently accumulated on such surfaces. Absence of paleosols and caliche suggests that any one surface was not exposed for long, although the ubiquitous iron oxide pigment demonstrates that oxygen was abundant in the diagenetic environment. The lack of ripplemarks and cut-and-fill structures argues against the fluviatile and tidal-flat origins often postulated for redbed sequences.

Perhaps these peculiar red siltstones are best understood as sediment transported by wind from an arid landscape to the east and deposited on a surface typically, but not invariably, covered by a shallow sheet of water (Renner, 1988). In this scenario, minor falls in sea level occasionally exposed broad areas of silt. At times of inundation, the silt-choked water was conducive to neither animal nor plant life. Carbonate deposition took place only when unusually high sea stands brought clear-water conditions.

To those who yearn for paleontological insight concerning the very last moment of the Paleozoic, it was a cruel trick of nature that recorded Wyoming's most complete record across the erathem boundary in the barren Goose Egg Formation, rather than in the fossiliferous sequence farther west. Not to say that deposition was literally continuous in eastern Wyoming, but no evidence, subsurface or surface, has been presented for a lengthy hiatus in aggradation of silt during the passage from Paleozoic to Mesozoic eras.

By contrast, paleontologists have long maintained that the faunal sequence in the Phosphoria and Park City Formations is three bricks shy of a load, as the saying goes, relative to more complete Upper Permian sections in the Tethyan region. Six biostratigraphic zones, based on conodonts and brachiopods, have been documented in the Phosphoria-Park City sequence of Wyoming. The uppermost one is thought to be no younger than earliest Capitanian in the West Texas standard section. Furthermore, this zone has been identified only in west-central Wyoming. Farther west, the fact that uppermost Permian rocks bear representatives of older zones supports the lithostratigraphic evidence there for a significant gap in the record at the erathem boundary.

The absence of fusulinids from the Phosphoria-Park City rocks comes as an unpleasant surprise to those accustomed to using them for biostratigraphic

work in this part of the Permian System. Their disappearance from this region is all the more mysterious considering their presence through much of Pennsylvanian and earliest Permian time. The best guess is that they were denied access to the Wyoming shelf from the west by low temperature waters of the upwelling system associated with the Phosphoria basin, and from the east by hypersaline conditions.

Unlike the fusulinids, several normal-marine groups continued to inhabit Wyoming while the Phosphoria-Park City sequence accumulated. Bryozoans, brachiopods, and crinoids are common constituents of all the carbonate members and can be found nearly to the top of the Park City Formation where it is overlain by Triassic mudstones. The highest beds of the upper member, however, show evidence of regression. At many localities, thin layers of laminated, unfossiliferous carbonate cap the member (**Figure 14**). Desiccation cracks are rare, but the upper surface commonly exhibits an alteration rind, silicified in places, suggestive of an unusual interval of exposure before sedimentation resumed with deposition of Triassic silt and clay.

Relief on this surface is typically on the order of inches rather than feet, and it is not overlain by a bed of Triassic conglomerate. The absence of a prominent disconformity contrasts sharply with the paleontological evidence of missing record and provides a final paradox for those who try to reconstruct the Paleozoic history of Wyoming.

Many valuable works dealing with Wyoming Permian rocks are referenced in important reviews by Peterson (1984) and Wardlaw and Collinson (1986). A regional synthesis by Rascoe and Baars (1972) provides perspective and is notable for illustrating subsurface relationships between the Permian of eastern Wyoming and adjacent states. Renner's (1988) study of the Permian-Triassic boundary beds is a fine example of integration of surface and subsurface evidence.

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Figure 14. Thin carbonate beds of Park City Formation (Permian) overlain by drab, slope-forming mudstones of Dinwoody Formation (Lower Triassic), northwestern end of Wind River Canyon.

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Mesozoic history of Wyoming

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Abstract

In the Late Triassic, Early Jurassic, and for much of the Late Jurassic, the Wyoming shelf was emergent, a terrestrial plain covered by lakes and streams and sand seas. For most of the rest of the early Mesozoic, shallow seas covered the shelf. The Early Triassic sea (Griesbachian) transgressed quickly to the east, somewhat beyond the former basin of the Permian Phosphoria Formation, and covered the Wyoming shelf. Before the end of the Griesbachian, terrigenous silty sediments began to prograde west over the shelf. During deposition of the overlying Red Peak Formation (Lower Triassic), a complex current system existed in the shallow sea and tidal plain. Red Peak facies are characterized by their red color, high silt and very-fine-sand content, abundant and diverse sedimentary structures, gypsum, and the scarcity of fossils. Above the Red Peak, the Alcova Limestone is a marine tongue extending eastward well out onto the shelf from the sandstone and limestone unit of the upper Thaynes Formation of southwestern Wyoming and southeastern Idaho. The overlying Crow Mountain Sandstone (Middle? Triassic) accumulated during a westward regression of the shelf sea. A tidal flat, beach, and marsh setting prevailed during deposition of the upper Crow Mountain. An unconformity separates the Crow Mountain and the Upper Triassic unnamed red-bed unit (central and western Wyoming) and the fluvial Jelm Formation (southeastern Wyoming). The Jelm (or unnamed red-bed unit) was laid down on a westward-prograding fluvial-deltaic plain in central Wyoming. Farther west, the interval was probably deposited in a shallow marine to paralic setting. The Popo Agie Formation, the youngest Triassic strata, was deposited in a vast river and lake system; volcanoes rained ash over most of western Wyoming and on parts of Idaho, Utah, Colorado and, likely, on adjacent areas in the western United States.

Dominantly eolian strata of Nugget Sandstone (Lower ? Jurassic) unconformably overlie the Popo Agie Formation, but contain beds that were laid down in several environments–eolian, small lakes within dune areas, inland sabkhas, and streams. The lower evaporite and red fine-grained clastic unit of the marine Gypsum Spring Formation (Middle Jurassic) lies on a prominent unconformity above the Nugget Sandstone. Evaporites and associated beds accumulated on sabkhas in arid and semi-arid climates. Before advance of the Middle Jurassic Sundance Formation seas, the stable Wyoming shelf was part of a regional low-lying erosion surface. The lower (Callovian) and upper (Oxfordian) Sundance divisions each include a major transgressional/regressional marine sequence. The lower Sundance sea is represented by the Canyon Springs Sandstone and the lower part of the Stockade Beaver Shale, the prograding retreat by the upper Stockade Beaver Shale, the Hulett Sandstone, and the Lak Red Beds. During the Late Jurassic, the Wyoming shelf had a paleolatitude between 35° and 39°N. The climate was warm and dry–winds blew from west to east over the Western Interior. Sundance seas were succeeded by terrestrial environments represented in the Morrison Formation (Upper Jurassic) by alluvial fans, braided and meandering streams, and lakes. The climate was strongly seasonal with high rainfall interrupted by brief dry seasons.

By the Cretaceous a north-south mountain welt with attendant foredeep had formed in western North America. The foreland basin was invaded by an epicontinental sea that began as a long northern arm and a short southern arm that joined in Late Albian, connecting the proto-Gulf of Mexico with the Cretaceous circumpolar sea. Lower Cretaceous strata are 5500 feet (1650 m) thick in the west, but less than 500 feet (150 m) over nearly all of the rest of Wyoming. Only in the westernmost part of the state, where alluvial fan and lake deposits occur, is there a record of

Neocomian time. In the early Aptian, streams carried gravel eastward, the first Cretaceous deposits over much of the state. During the Late Albian, seas spread over Wyoming, beginning the characteristic transgressions and regressions that lasted until the mid-Maastrichtian. Upper Cretaceous beds are 18,000 feet (5400 m) thick where the western foredeep persisted, 16,000 feet (4800 m) thick in south-central Wyoming, and less than 4000 feet (1200 m) thick in the central western part. While the Late Cretaceous transgressions and regressions were dominant, a north-south belt of complexly interfingering shallow marine, deltaic, barrier-island, and coastal-plain environments existed across Wyoming with predominately marine deposits accumulating on the east and nonmarine and marginal-marine deposits on the west.

Introduction

The Mesozoic Era ("time of middle life"), from about 245 to 65 Ma (Mega-annum or million years ago), is noteworthy because of the dinosaurs' success, because Cretaceous strata contain the world's second largest coal reserves and a large part of its petroleum, and because of the stunning landscapes its rocks provide–red beds, chalks, gray-and-black claystone badlands. Eldredge (1991) emphasizes there is a true intermediacy in the "archaic aspect of the Paleozoic, and the modern, or very nearly so, appearance of fossils of the Cenozoic."

Pangaea broke up during the Mesozoic, profoundly affecting the climate and the evolution of life, particularly that of nonmarine plants and vertebrates. Major mass extinctions bracket the Mesozoic, and a mass extinction ended the Triassic Period. Thus, three of the "Big Five" extinctions of species began, fell within, and ended the era.

Beds of the Lower Triassic Red Peak Formation–the red-bed sequence of the Chugwater Group–are perhaps the most striking sedimentary strata in Wyoming. Wherever they occur in long bands–on the northeast flank of the Wind River Range (**Figure 1**), on the enclosing rim of the Bighorn Basin, in the thrust sheets of the fold-and-thrust belt, in Jackson Hole–they glow with the remembered warmth of the Triassic seas and tidal plains.

Travelers affected by rocks will remember forever this red-bed land. Of all outcrops red beds are probably the most spectacular rocks to tourists and geologists alike. The widespread, notable, red bands of this region are the Permian-Triassic Goose Egg Formation, the Triassic Red Peak and Jelm Formations, and the Jurassic lower Nugget and Gypsum Spring strata. After the colorful Triassic and Jurassic rocks, the blacks and grays and yellowish-browns of Cretaceous rocks are somber. The Cretaceous is early Van Gogh, the early Mesozoic, late. Tramping through Late Cretaceous Mancos mudstone, one may long for what seems to have been, in retrospect, the romantic age of the early Mesozoic.

Triassic rocks in Wyoming belong to the Dinwoody Formation (older) and the Chugwater Group (younger). From oldest to youngest, Chugwater includes: Red Peak Formation (Lower Triassic), Alcova Limestone, Crow Mountain Sandstone (Middle? Triassic), Jelm Formation (Upper Triassic), and Popo Agie (pronounced Pō-pō-sāh) Formation (Upper Triassic). The Jurassic rocks



Figure 1. Part of Triassic and Jurassic sequence exposed on northeastern flank of the Wind River Range. View to the northwest from State Highway 28. Red Peak Formation (hogback cliffs in middle ground), Alcova Limestone (A), Crow Mountain Sandstone, Jelm Formation, Popo Agie Formation, and Nugget Sandstone (N).

are: Nugget Sandstone (Lower? Jurassic), Gypsum Spring Formation (Middle Jurassic), Sundance Formation (Middle Jurassic), and Morrison Formation (Upper Jurassic). The fold-and-thrust belt bears different names, to be identified at the appropriate places. The dark gray and black claystone and yellowish-brown sandstone, characteristic of Cretaceous rocks, contrast sharply with the bright-hued, early Mesozoic rocks below. Cretaceous stratigraphic nomenclature, too extensive to summarize here, is given at appropriate places in the text and in the stratigraphic chart by Love and others (1993). **Table 1** includes a stratigraphic column and other information for the early Mesozoic. Place names are shown on **Figure 2**.

Table 1. Stratigraphic sequence, depositional environments, and sandstone types of early Mesozoic strata, Wyoming.

Age	Formation	Depositional environment[1]	Rock type[2]	Sandstone and siltstone type; QFL[3] (ss, sltst)	Thickness (feet)
Jurassic	Morrison	A, L	F, Ss, Cgl, Carb, V,R	quartzarenite, sublitharenite	100-500
	Sundance				
	Windy Hill	M	Ss, F	quartzarenite, sublitharenite	5-57
	Redwater	M(td)	F, Ss, Carb		0-135
	Pine Butte	M	Ss, F		0-35
	Lak	E, M(s)	Ss, F, Ev, R	quartzarenite, sublitharenite	0-75
	Hulett	M(wt)	Ss, F	quartzarenite, sublitharenite	3-100
	Stockade Beaver	M(o)	F		0-95
	Canyon Springs	M(td, cs)	Ss, F, Carb	quartzarenite, sublitharenite	1-65
	Gypsum Spring	M(s)	F, Ev, Carb, R		0-250
	Nugget	E, L	Ss, F, Carb, Ev, R	82-14-4, 74-24-2	0-800
Triassic	Jelm	A, L, M(s)	Ss, F, Carb, R		25-650
	Popo Agie	L, A	F, Ss, Cgl, V, R		0-275
	Crow Mountain	M(td)	Ss, F, R	85-13-2, 79-16-5	0-175
	Alcova	M	Carb, Ev	77-18-5	0-25
	Red Peak	M(td)	F, Ss, Carb, Ev, R	73-24-3, 60-31-9	400-1250
	Dinwoody	M	F, Ss, Carb		15-250

[1] Abbreviations: A = alluvial; E = eolian; L = lacustrine; M = marine; s = shallow; td = tide dominated; wt = wave and tide dominated; cs = clastic shoreline; o = offshore

[2] Abbreviations: Cgl = conglomerate; Ss = sandstone; F = siltstone and claystone; Carb = carbonate rocks; Ev = evaporites; R = redbeds or varicolored rocks; V = volcanic or volcaniclastic rocks

[3] Percentages of quartz, feldspar, and lithic fragments

Triassic Period

Dinwoody Formation

By the latest Paleozoic all of the continents were assembled to form the universal megacontinent Pangaea. Rifting with an outpouring of basalt about 200 Ma (early in the Jurassic Period), signaled the breakup of the supercontinent. By the end of the Jurassic, breakup had been underway for 65 my (million years).

Unfortunately, as in most of the rest of the world, the sequence of strata in Wyoming from the Upper Permian into the Lower Triassic is incomplete. In central and eastern Wyoming, the discordance between Permian and Triassic rocks is generally slight or not apparent; the rocks above and below the boundary are very similar; and fossils are rare or absent (Love, 1948; Kummel, 1954; Burk, 1956). Boyd and Maughan (1973) concluded that deposition in western Wyoming was continuous during the latest Permian and earliest Triassic and placed the era boundary at the base of the Dinwoody Formation. Where the Permian-Triassic boundary occurs within a red-bed sequence in southeastern Wyoming, erosional surfaces are present within the red beds, but which surface, if any, marks the boundary is impossible to say.

The last widespread episode of marine deposition in the Cordilleran miogeocline that followed a depositional and structural pattern inherited from the late

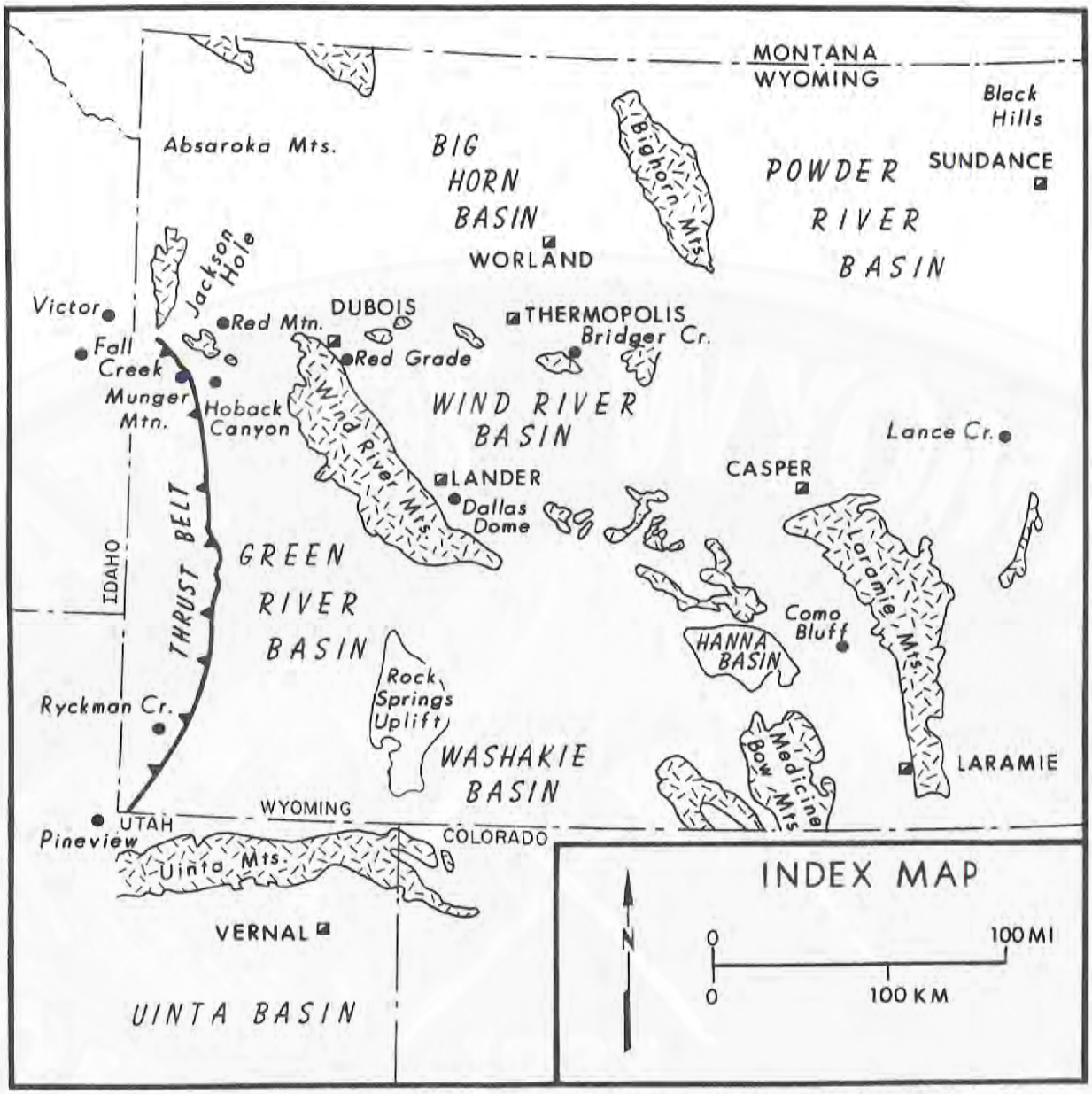


Figure 2. Index map showing Precambrian outcrops, ranges, basins, and place names referred to in the text.

Paleozoic, and possibly from the late Precambrian (Carr and Paull, 1983), is recorded in Lower Triassic rocks (**Figure 3**). Early Triassic strata register the end of marine deposition along the passive continental margin of western North America. Speed (1978, 1979) suggested that the great range in depositional and tectonic patterns in the eastern Cordillera during the Early Triassic was related to the accretion of a Permian oceanic-arc terrane onto the western margin of North America.

The Dinwoody Formation occupies about the same geographic area as the Phosphoria Formation–western Wyoming, southwestern Montana, southeastern Idaho, and northern Utah. The basin's center was in southeastern Idaho where the Dinwoody reaches a maximum thickness of 2400 feet (720 m) and includes strata both older and younger than those in the type locality in the Wind River Range near Dubois, Wyoming.

Siltstone is the principal rock in the Dinwoody Formation. It is light-colored–gray or green or light brown and, in some places, orange–sandy, and streaked and interbedded with anhydrite in the subsurface and gypsum at the outcrop.

The masters of the Dinwoody seas in western Wyoming were gastropods and bivalves, but they were much less diverse than those living on the outer shelf in what is now southeasternmost Idaho. To the end of the Permian, there were few gastropods to be seen at all and "Paleozoic-like snails are simply more characteristic of the Triassic than they are of the Upper Permian," said Eldredge (1991). In the limey siltstone of western Wyoming, horizontal-paired trails and less

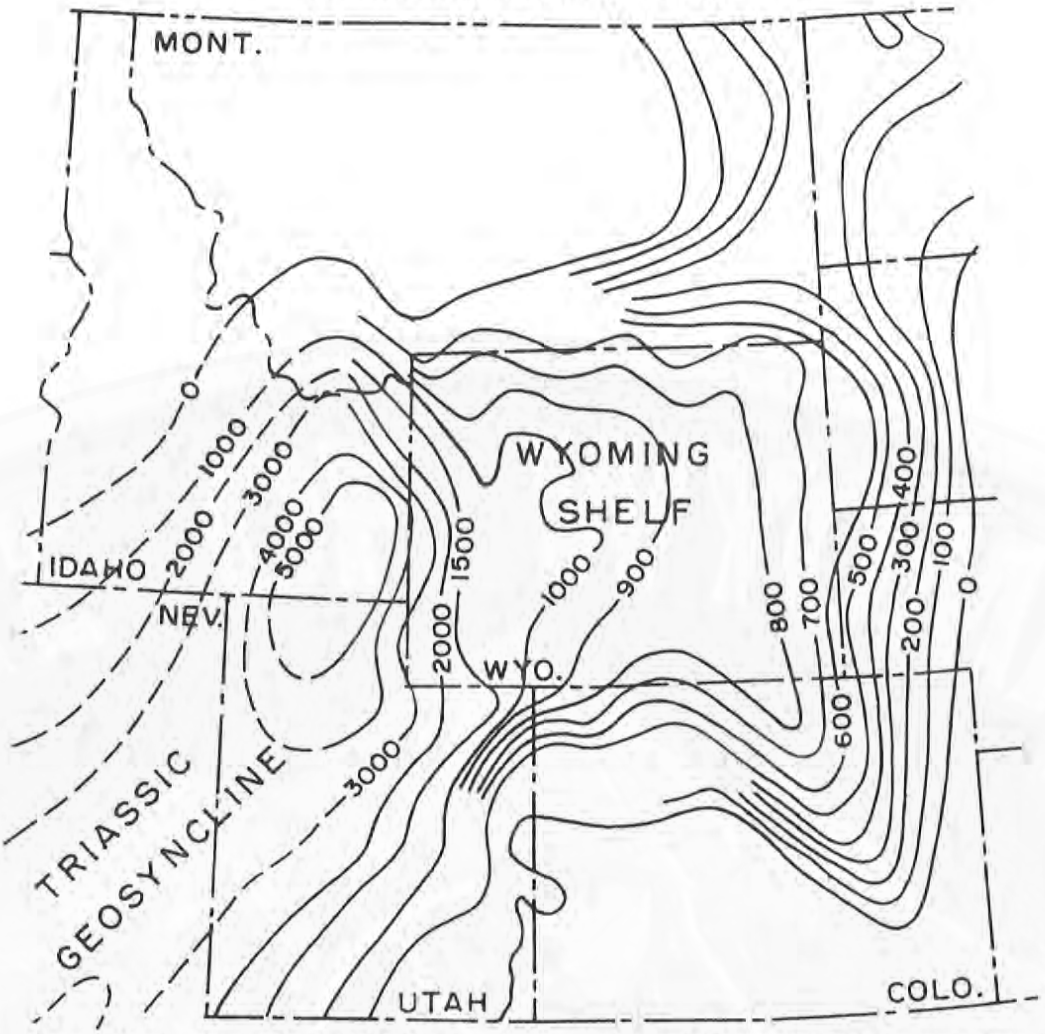


Figure 3. Regional paleogeography of Western Interior during Early Triassic (after McKee and others, 1959). Isopach lines in feet; isopachs in western Wyoming include Dinwoody and Red Peak Formations.

common U-shaped burrows of the *Cruziana* shallow-marine association are well preserved (Carr and Paull, 1983).

The Early Triassic sea (Griesbachian Stage) transgressed rapidly to the east, somewhat beyond the former Phosphoria basin, and covered the Wyoming shelf. Griesbachian deposits of the Dinwoody Formation are apparent as far east as central Wyoming (Carr and Paull, 1983). In southeastern Wyoming, the Freezeout Shale and the Little Medicine Members of the Goose Egg Formation—Permian-Triassic red beds, carbonates, and evaporites—grade westward into yellowish calcareous siltstone in the lower part of the Dinwoody Formation (Maughan, 1964). Across the Wyoming shelf the basin gradient was low. Thus, slight variations in sediment supply or a minor change in relative sea level affected broad areas (McKee and others, 1959; Collinson and Hasenmueller, 1978).

Before the end of the Griesbachian, terrigenous silty sediments began to prograde westward over the Wyoming shelf (Carr and Paull, 1983). This progressive regression with minor fluctuations continued throughout the Dienerian (McKee and others, 1959).

Red Peak Formation

The most notable feature of the Early Triassic Red Peak Formation is its red color. When the Chugwater Group is labeled a red-bed sequence, it is the Red Peak Formation that has impressed he observer. The red pigmentation occurred during diagenesis (Picard, 1964, 1965, 1966). The sediment was not red, and terrigenous grains were not coated with hematite at deposition.

In Wyoming, the Red Peak Formation contains five recognizable facies that are, from base to top: silty claystone, lower platy, alternating, upper platy, and sandy (**Figure 4**; Picard, 1967, 1978). Especially notable is the general upward increase in abundance of coarser grains and the general upward increase in



Figure 4. Upper part of Red Peak Formation at Red Mountain northeast of Jackson, showing interval containing alternating facies of Red Peak, upper platy facies, variegated sandy facies, Alcova Limestone, and Jelm Formation.

number and thickness of coarser beds. The thin upper platy facies (about 20 to 75 feet thick) reflects a return to the depositional conditions of the lower platy facies.

From the character, abundance, and associations of bedding types and sedimentary structures, it is possible to distinguish the principal depositional environments. For each facies, these are: silty claystone=transitional paralic; lower platy=tidal flat; alternating=tidal flat and shelf; upper platy=tidal flat; and variegated sandy=nearshore marine.

From the thin (1 to 2 feet thick) gypsum beds–especially in the lower platy and alternating facies–casts of salt cubes, mud cracks, raindrop impressions, and scarcity of fossils in the Red Peak Formation, it is inferred that the climate was warm to hot and semiarid to arid during its deposition. Oxidizing conditions prevailed. The depositional basin was within 30° of the Early Triassic equator. In the dry climate, there was little vegetation and a low water table, which facilitated formation and preservation of ferric oxides in the red beds.

During deposition of the Red Peak Formation, a complex current system existed in the shallow sea over the Wyoming shelf. Its waters moved in two major directions–to the northeast and northwest–with a minor southwest-directed current. The three major Red Peak current directions apparently record onshore, offshore, and longshore movement produced by wave drift, rip, and longshore currents, respectively. Choosing between two possible reconstructions that would satisfy the inferred current system, Picard and High (1968) suggested that the average shoreline trended northwest-southeast. A northwest-southeast trending shoreline agrees closely with thickness trends of the Dinwoody and Red Peak Formations in western Wyoming and with a detailed thickness map of the Red Peak.

If one judges the quality of life during Red Peak time by the fossils found in these Early Triassic red beds, overcrowding was not a problem. The Red Peak contains small burrows, but they are not definitive as to maker or environment and have not been studied in any detail. Lull (1942) described small reptile footprints from strata about 35 feet (11 m) below the Alcova Limestone in the upper platy facies in central Wyoming. Love (1948) reported a single pelecypod from the northern Wind River Basin. Picard (1967) found fish scales in the alternating facies in the southern Bighorn Basin. At three Red Peak localities representing different stratigraphic horizons, Boyd and Loope (1984) described distinctive sole marks that they interpreted as scratches made by the toes of amphibious tetrapods.

These indications of Red Peak life are all that is known–small burrows, small reptile footprints, a pelecypod, fewer fish scales than one would see near any fishing hole, and indications that a mysterious small group of amphibious tetrapods gained geological immortality by scratching in the mud in several places tens of miles apart. Speculation about these red-bed depositional environments thus rests largely on the physical evidence in the rocks and not on the biologic evidence, since diagnostic fossils do not exist.

The red beds intertongue in western Wyoming and eastern Idaho with limestone, dolomite, sandstone, and siltstone of the Thaynes Formation (Picard and others, 1969). These rocks preserve for us, even in the dolomitized intervals, pelecypods and crinoids, ammonoids, and, less abundantly, brachiopods, gastropods, cephalopods, bryozoans, foraminifers, and algae. At Hoback Canyon in western Wyoming, where the sequence is still dominantly red siltstone and sandstone, there are rare invertebrates–pelecypods especially–in the carbonate tongues and red-bed intervals.

Abundant gypsum and dolomite in the upper part of the sections at Red Mountain (**Figure 4**), Hoback Canyon, and Munger Mountain indicate deposition in a restricted marine environment. Similar environments probably prevailed also during deposition of the lower part of the Red Peak-Thaynes sequence at the same localities. To the west, the Thaynes is more calcareous, and only a few feet of dolomite occurs at Fall Creek in eastern Idaho. A series of carbonate banks on the shelf edge in eastern Idaho would have produced restricted conditions in the shallow water among and just behind the banks (**Figure 5**). Sediment accumulating in this area therefore would contain hypersaline pore solutions. The resulting penecontemporaneous dolomitization, associated with minor bedded gypsum, is reflected in the facies relations between western Wyoming (Red Mountain) and easternmost Idaho (Victor). On the east, the wide Wyoming shelf was largely unaffected by the shelf-margin banks. Water circulation was generally unrestricted.

As noted in western and west-central Wyoming, there is a conspicuous change in rock types at the Dinwoody-Red Peak (silty claystone facies) contact. Poorly sorted, fine-grained material lies above the carbonate, evaporite, and siltstone strata of the Dinwoody. Though there is a general upward-coarsening, deposition of similar fine-grained material continued nearly to the end of Red Peak deposition. In the Red Peak Formation, silt is the dominant clastic constituent of the rocks, followed by very-fine sand, and then clay-size particles.

Large amounts of wind-blown silt apparently fell on the Wyoming shelf, certainly during Red Peak deposition, but perhaps in the Late Permian, continuing through Dinwoody and Red Peak deposition. The last Dinwoody sea was retreating slowly and was very shallow or partly ponded when lowermost Red Peak grains began to accumulate. The Wyoming shelf then had a paleolatitude between about 15° and 25°N (Robinson, 1971, 1973). Silt rained down on the land adjoining the sea, on islands, and in the shallow sea from great dust storms carried by northeast trade winds. Waves and currents in the remnant sea sorted the grains and formed sedimentary structures and bedding.

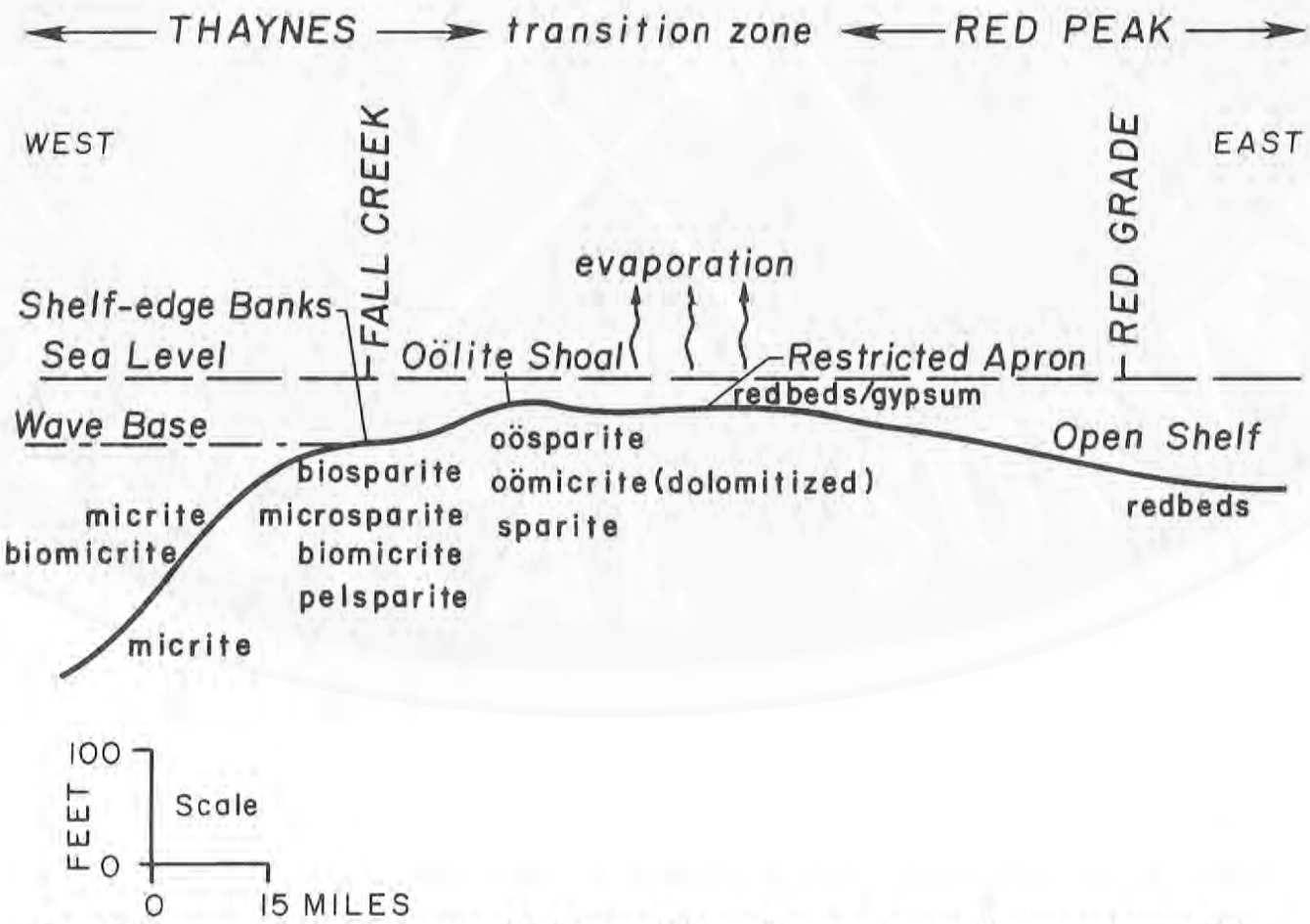


Figure 5. Paleogeography during deposition of Red Peak and Thaynes Formations, western Wyoming and eastern Idaho.

The end of typical Red Peak deposition came with a flood from the east of very-fine sand over well-sorted and poorly-sorted silt of the upper platy facies. The variegated sandy facies may be unconformable on the upper platy facies. The time that followed was one of sand and carbonate deposits cemented in a shallow sea and tidal flat across the Wyoming shelf. The youngest Red Peak strata–the sandy facies–show great variation in color–grayish red, pale red, pale olive, pale green, dark yellowish orange–where the Alcova Limestone overlies the sandy facies, and are dominantly red at localities where the Alcova Limestone is absent. The variegated nature of the sandy facies is related to reduction of red iron oxides by water expelled from the Alcova during diagenesis.

Alcova Limestone

The Alcova Limestone (**Figure 6**) is the most widespread and easily recognized marker bed in the early Mesozoic strata of Wyoming, and certainly one of the most widespread marker beds in North America, covering at least 50,000 square miles (130,000 km^2). It is an incredibly thin formation to blanket such a large area–less than a foot thick in places. During Alcova deposition the Wyoming shelf was stable, one of the stillest places and times of the Mesozoic era.

The Alcova sea transgressed quickly over the Wyoming shelf to east-central Wyoming and, after leaving what is now up to about 25 feet (7.5 m) of carbonate rock, regressed with dispatch westward toward the continental margin in westernmost Wyoming and adjacent Idaho.

Though the Alcova is the best developed carbonate tongue that extends out of the basinal carbonate rocks well into the Triassic red-bed facies, it is difficult to know with what basinal rocks it is to be correlated. Distances between outcrops are great, in some places tens of miles. Wells are scarce in the critical parts of Wyoming and Idaho where they are necessary for confident correlation of the buried Alcova. In western Wyoming and eastern Idaho, structural complexity makes correlation difficult. The precise correlation of the Alcova Limestone with the southeastern Idaho section is thus especially perplexing to geologists (see table 1 in Picard and others, 1969, and in Picard, 1993). In 1969, Picard and others suggested that the Alcova is a marine tongue extending eastward onto the Wyoming shelf from the sandstone and limestone unit of the upper Thaynes Formation (Kummel, 1954). The exact position of the Alcova equivalent in the "sandstone and limestone unit," which is about 150 feet (45 m) thick at Hoback Canyon, was not determined. In fact, there are several thin carbonate layers within about 90 feet (27 m) of dolomite, gypsum, and fine-grained terrigenous red beds at Red Mountain, which is about 53 miles (85 km) west of the well-studied section at Red Grade, southeast of Dubois. At Red Grade there is only a single thin carbonate unit in the red-bed facies, the Alcova Limestone.

In the Bighorn Basin, the Alcova Limestone occurs along the southern margin, but it is absent to the north by truncation or by local facies change to sandstone. The Alcova is locally missing in the northwestern part of the Wind River Basin, because in this area there were islands in the Alcova sea. The interval is represented by sandstone, part of the Crow Mountain Sandstone.

The Alcova sea may have been full of life, but the remains are slight, restricted, and undistinctive (Carini, 1964). The undistinguished Alcova biota includes molluscan, reptilian, and algal components.

> *The molluscan assemblage comprises two genera of pelecypods similar in external configuration and ornamentation to the freshwater forms* Naiadites *and* Sphaerium, *and one genus of gastropod (*non Natica*) possibly allied to the basommatophorids,*

said Carini (1964). Case (1936) described Corosaurus alcovensis, a semi-aquatic reptile, from the Alcova Limestone. Later on, von Huene (1948) redefined the reptile, considering it to be an ancestral plesiosaur. Carini noted an undescribed thecodont reptile, a phytosaur or Coelophysis-like dinosaur.

The Alcova flora has given character to the rocks to a greater extent than the few molluscs and reptiles. Stromatolitic algae, inferred to be the work of blue-green or green algae, is spectacularly developed lo-

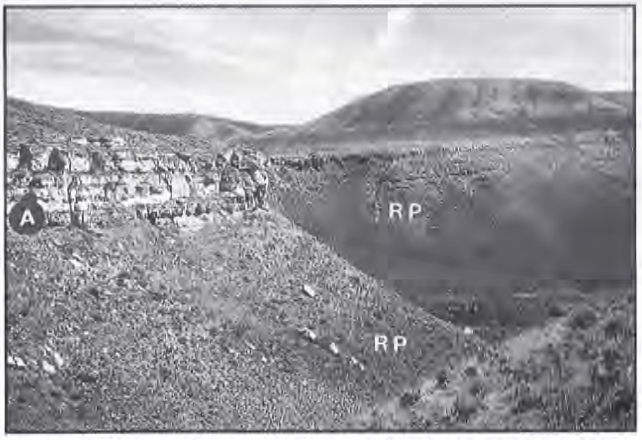


Figure 6. Slope and cliff of Red Peak Formation (RP), variegated sandy facies, Alcova Limestone (A), and Crow Mountain Sandstone (light-colored sandstone above the Alcova in foreground left) at Bridger Creek.

cally, especially in the lower Alcova. Algae is the most volumetrically significant organic constituent in the formation.

The Alcova Limestone is the deposit of a "desalted isolated lake-sea comparable to the extant Caspian Sea or Sea of Aral," said Carini (1964). When he wrote, few criteria existed for recognizing ancient lacustrine strata, and he cited little actual evidence of an inland sea or lake. In his view, the lack of saline-marine fossils meant that Alcova waters were fresh to brackish. Limited stable isotope values from Alcova algal facies lend little support to Carini's lake hypothesis. $\delta\ ^{13}C$ values of three samples are -1.0133 (the mean); the $\delta\ ^{18}O$ values are -5.2203 (Cavaroc and Flores, 1991). Cavaroc and Flores seemed to suggest that these results indicate brackish water conditions. An alternative interpretation would be that the Alcova $\delta\ ^{13}C$ values are far from indicating fresh or brackish water, but they are close to typical chalks. All three $\delta\ ^{18}O$ values are typical of marine limestone. The Ca:Mg ratios are greater than 70. Nothing particularly unusual can be said of the few microprobe and stable isotope measurements other than that they are compatible with the interpretation of a marine origin for the Alcova Limestone.

Pipiringos and O'Sullivan (1978) postulated an inconspicuous but widespread unconformity (their 'R-2') separating the Alcova Limestone and the Crow Mountain Sandstone. Its duration is unknown. They recognize it "at the base of the Crow Mountain Sandstone in north-central Wyoming and at the base of the Jelm Formation in Wyoming and north-central Colorado." Pebbles of Alcova Limestone occur in the basal few inches of the Crow Mountain Sandstone and in the Jelm Formation (Love and others, 1945b; Love and others, 1947; Tohill and Picard, 1966; Pipiringos, 1968).

Crow Mountain Sandstone

Crow Mountain is the coarsest Triassic sandstone band in Wyoming (**Figure 6**). The grand hogbacks are held up by the sandy facies, the thin resistant Alcova Limestone, and the lower part of the basal Crow Mountain Sandstone.

The only common indications of ancient life in the Crow Mountain are trace fossils in the sandstone and siltstone. Trails that parallel bedding planes in cross-stratified sandstone are notable. Some specimens contain an internal structure of meniscate backfill. There are small vertical, inclined, and horizontal burrows. The burrowing organisms continued to churn sediments in some intervals, so that all of the original stratification was destroyed.

In northwestern Wyoming, currents in the Crow Mountain sea apparently moved onshore to the northeast and offshore to the southwest (Tohill and Picard, 1966). At many localities, a significant number of paleocurrent directions are discernible at approximate right angles to the predominant northeast or southwest directions. These probably reflect currents moving along shore to the northwest or southeast. The orientation of the average shoreline is interpreted to have been northwest-southeast, at least in what is now northwestern Wyoming.

Cross-stratification characterizes the basal sandstone unit of the Crow Mountain, which is red, reddish-brown, yellowish-red, and brown calcareous sandstone with well-sorted, subrounded, very fine grains. Fine and medium grains occur, but in smaller amounts. The grains have been recycled, the majority of them several times. The medium-and large-scale trough (dominant) and planar cross-stratification signifies deposition in upper parts of the lower-flow regime. Current action in the shallow sea was wave and tidal driven.

Wave-formed ripple marks, which indicate formation in the lower part of the lower-flow regime, characterize the upper sandstone and siltstone unit. Red and reddish-brown and drab (non-red), calcareous, very-fine-grained sandstone interbedded with red and drab calcareous siltstone distinguish the sequence. The upper unit includes less porous and permeable layers than lower Crow Mountain Sandstone. Depositional breaks are common. Sediments were periodically exposed as indicated by the frequency of shrinkage cracks and claystone-pebble conglomerate. We have suggested a tidal flat, beach, and marsh setting during upper sandstone and siltstone deposition (Tohill and Picard, 1966).

The Crow Mountain apparently accumulated during a westward regression of the shelf sea. At the beginning of sandy facies deposition, regional uplift of the basin and source areas on the north, east, and south occurred. The basal sandstone thickens toward the north and east, and the clastic strata are coarser grained on the north and east.

In the sixties, above the Crow Mountain in the Wind River Basin and in the southern Bighorn Basin, Tohill and Picard (1966) recognized a red-bed unit–sandstone dominantly, but also siltstone, claystone, claystone-pebble conglomerate–to which they gave no formal name. An unconformity separates the unnamed red-bed unit and the Crow Mountain. The Crow Mountain is Middle (?) Triassic, the overlying red-bed interval, probably Upper Triassic. Pipiringos and O'Sullivan (1978) have called the unconformity

'Tr-3', and said it "is one of the most widespread, conspicuous, and widely recognized unconformities" in the Triassic and Jurassic in the Western Interior United States. In northeastern Utah, it lies between the Gartra Formation and the Moenkopi Formation (McCormick and Picard, 1969). Pipiringos and O'Sullivan (1978), suggesting that the Tr-3 lasted less than a million years, placed the unconformity within the Crow Mountain Sandstone in the Wind River and Bighorn Basins, but apparently they defined the Crow Mountain there differently than we did (Tohill and Picard, 1966).

Jelm Formation

During deposition of the Crow Mountain Sandstone, much of the Wyoming shelf was flooded by a shallow sea. Streams emptying into it in southeastern Wyoming deposited the lower Jelm Formation. A general withdrawal followed, during which the fluvial and fluvial-lacustrine sediment of the Jelm and Popo Agie Formations prograded over the shallow marine Crow Mountain Sandstone.

In 1917, Knight separated the Jelm Formation from the Chugwater Formation in southeastern Wyoming because it is unconformable on the Chugwater (Red Peak Formation). In southeastern Wyoming, the upper 250 feet (75 m) of the red-bed sequence includes a bed of bone- and plant-bearing conglomerate containing a Triassic (?) vertebrate fauna. Inasmuch as the Alcova does not extend that far east, exact placement of the Chugwater-Jelm contact is unknown relative to the Chugwater of central Wyoming. It lies at the approximate position of the Alcova, however, and the Jelm in the type area is thus approximately equivalent to the Crow Mountain and Popo Agie of central and western Wyoming described by previous workers.

Branson (1927) believed the Jelm and Popo Agie deposits to be representatives of local, short-duration environments that probably were not contemporaneous. Hubbell (1954) used Jelm and Popo Agie interchangeably. Pipiringos (1968) considered the Jelm equivalent to the Crow Mountain of central Wyoming. High and Picard (1969) interpreted the Jelm Formation to be a facies equivalent of the upper Crow Mountain Sandstone in southeastern Wyoming, and to unconformably overlie the Crow Mountain throughout central Wyoming.

The central Wyoming Jelm (or unnamed red-bed unit) differs from more eastern sections in the rarity of conglomerate and the presence of clayey limestone and platy, ripple-marked, carbonate-cemented sandstone. Cross-stratified, channel-filling, red sandy siltstone and sandstone remain the dominant rocks. They were deposited on a generally westward-prograding fluvial deltaic plain (High and Picard, 1969).

From their study of the Chugwater Group along the southwestern margin of the Powder River Basin, Cavaroc and Flores (1991) also interpreted the red beds as representatives of fluvial-deltaic plain deposits. They noted thick (as much as 35 feet) lenticular sandstone, and "poorly-sorted channel fills and tabular splays" (generally <10 feet thick). The cross-strata in the channels indicate lower-flow regime conditions, but the planar lamination represents deposition in the upper-flow regime. Channel-filling sandstone grades upward into ripple-dominated beds. Some sandstone shows abundant burrows.

High and Picard (1969) showed the Jelm tongue grading into the basal part of the Ankareh Formation (**Figure 4**) in western Wyoming. The interval consists of dark reddish-brown, platy, calcareous siltstone and very fine- to fine-grained sandstone with abundant small-scale cross-stratification and ripple marks. Likely, now we would assign these beds in western Wyoming to the unnamed red-bed unit, which Tohill and Picard (1966) used. These rocks closely resemble platy facies of the Red Peak Formation, which, on the basis of sedimentary structures and stratigraphic relationships, are interpreted to be tidal-flat deposits. Thus, in western Wyoming this sequence may be shallow marine to paralic.

Popo Agie Formation

The Popo Agie Formation was deposited in a vast river and lake system. During deposition, volcanoes rained ash over most of western Wyoming and on adjacent parts of what is now Idaho, Utah, and Colorado. In south-central, central, and western Wyoming, High and Picard (1965, 1969) recognized four easily distinguishable rock units: lower carbonate (oldest); purple; ocher; and upper carbonate (**Figures 7** and **8**). The upper three units also crop out in the Uinta Mountains area of Utah and Colorado, and extend westward to north-central Utah maintaining thicknesses comparable to those in correlative outcrops on the western end of the Uinta Mountains.

Lake Popo Agie was one of the largest (70,000 square miles +) of the known ancient lakes (Picard and High, 1972, 1985). Overlying strata truncate the Popo Agie Formation on the north and east. The lake thus may have been much larger than 70,000 square miles (182,000 km^2). On the west and south, lake beds of the Popo Agie Formation grade into river beds.



Figure 7. Purple, ocher, and upper carbonate units of Popo Agie Formation unconformably overlain by Nugget Sandstone southeast of Dubois on U.S. Highway 287.



Figure 8. Unconformity separating lower Nugget Sandstone (Jurassic) from Popo Agie Formation (Late Triassic) at Red Grade in the Wind River Basin. My sons are standing near the base of the Popo Agie.

The lower carbonate unit of the Popo Agie Formation is a carbonate pebble conglomerate deposited in a stream environment and associated local ponds and small lakes. The purple unit is a silty smectitic mudstone and claystone laid down on floodplains by low gradient steams. Local beds of analcimolite (sedimentary rock with 50% or more of the zeolite mineral analcime) indicate that isolated lakes existed (Keller, 1952). Location of volcanoes there at the time is unknown. Ash, washed in by streams and carried to the basin by winds, settled on the vast alluvial plain to become clay minerals (smectite). The ocher unit is silty, analcimic, smectitic mudstone and analcimolite. It represents a widespread and persistent lake. The volcanic source areas continued to be active during deposition of the ocher unit. The same general conditions continued during deposition of the upper carbonate unit, which is mostly clayey dolomite and dolomitic mudstone. Evidently volcanism ended, however, while the lake was still active, and deposition of analcimic smectitic mud was replaced by illitic dolomite in the upper carbonate unit.

The lack of bedding and sedimentary structures in the Popo Agie Formation is striking: it is similar in its few bedding types and primary structures to many shallow marine carbonate units. This contrasts sharply with the abundant and diverse bedding and sedimentary structures in the shallow marine, fine-grained, terrigenous clastics of the Red Peak Formation. Disturbed bedding and intraformational conglomerate characterize the Popo Agie. Horizontal stratification, cross-stratification, homogeneous (no apparent bedding) stratification, and sedimentary dikes occur rarely in the formation. The Popo Agie is burrowed. The abundance of homogeneous bedding, disturbed bedding, and burrows indicates that much original bedding has been destroyed.

On the basis of its fauna, the Popo Agie Formation is assigned to the Late Triassic when dinosaurs dominated land habitats and all of the present continents remained joined in the super continent Pangaea.

The Popo Agie vertebrate fauna consists of the phytosaurs *Paleorhinus* and *Angistorhinus*, the metoposaurid *Eupelor*, a pseudosuchian *Dolichobrachium*, the therapsids *Brachybrachium* and *Eubrachiosaurus*, a fresh water lungfish *Ceratodus*, and the dinosaurs *Poposaurus* and *Agialopus* (Colbert, 1957; Gregory, 1957). High and others (1969) found the anterior end of a vertebra–probably from about the last third of a reptile tail–with phytosaur remains in the Uinta Mountains area north of Vernal, Utah. Wann Langston, Jr. (personal communication, 1969) suggested that it was from *Trilophosaurus*, which is best known from the Upper Triassic Dockum Group of Texas.

In the Late Triassic, phytosaurs were among the most numerous reptiles, similar in habits and appearance to modern crocodiles. They were, as well, the ecological predecessors of these forms, although not in themselves directly ancestral.

The fresh water pelecypod *Unio* is the only reported invertebrate in the Popo Agie Formation (Berry, 1924). Coprolites occur. Fossil plants include a large *Equisetum*, cycadophytes, and many wood fragments (Berry, 1924).

Reconstruction of the paleogeography gives the broad outlines of Lake Popo Agie. A lake origin for much of the Popo Agie is inferred from paleontologic, mineralogic, geochemical, and regional stratigraphic considerations. The Popo Agie fauna is nonmarine. Although distinctions between river and lake environ-

ments usually are not definitive based on fossils, the presence in the Popo Agie of large predators such as phytosaurs and labyrinthodont amphibians indicates semi-permanent water bodies.

The abundant analcime and carbonate in the Popo Agie Formation also suggests lake origin rather than stream origin. Inasmuch as streams are commonly acidic, and calcite is unstable at pH values less than about 7.8, fresh-water limestone indicates a lake origin. Similarly, authigenic analcime requires conditions more basic and higher concentrations of alkalis than are usual in streams.

Jurassic Period

Nugget Sandstone

The supposed facies or intertonguing relationships between the lower Nugget Sandstone and the upper Popo Agie Formation are cited as evidence that the Nugget Sandstone may be Triassic rather than Jurassic in age (Love, 1957). However, in 1965, High and Picard reported evidence of an unconformity at the base of the Nugget Sandstone (**Figure 8**). Apparently, the discordance occurs throughout Wyoming and is present in Utah and northwestern Colorado. Locally, older beds are tilted and folded and the unconformity is an angular one. This is especially notable in the southern Uinta Mountains area where broad folds formed during pre-Nugget Sandstone warping.

In Wyoming and Utah, marked change is apparent in the rock types and mineral content between the analcimic rocks of the purple and ocher units of the Popo Agie and the sandstone and siltstone of the lower Nugget Formation. Illite is the characteristic clay mineral in the Nugget; smectite in the underlying Popo Agie. Analcime does not occur in the lower Nugget except in the ocher granules and pebbles of the basal conglomerate. Ocher, light green, and other variously-colored granules and pebbles (up to 0.5 inches in diameter) in the conglomerate are the result of erosion of the Popo Agie, but were not transported far.

Crustaceans, molluscs (fresh-water bivalves), reptiles, fragmentary plant remains, scorpion tracks, and organic trails of unknown origin are present in the Nugget Sandstone and in the probably equivalent Navajo Sandstone south of the Uinta Mountains in Utah and in northern Arizona (Picard, 1977). Fossils are rare in the Nugget, however, as is the case in ancient wind deposits throughout the world. The Nugget and Navajo Formations are much studied, but no one has found them to contain marine fossils.

The Nugget Sandstone in Wyoming exhibits less evidence of ancient life than is the case on the south flank of the Uinta Mountains, in the north-central Wasatch Mountains, and in equivalent strata farther south (Picard, 1993). Nugget Sandstone in the Wind River Range and Uinta Mountains is divided into two informal rock units, an older thinly bedded facies and a younger cross-stratified facies (**Figure 9**). The thickness of the formation varies from less than 100 feet (30 m) to about 1000 feet (300 m). The thinly bedded facies ranges from about 25 to 125 feet (7.5 to 37.5 m) thick in the subsurface. It is a variable sequence of clayey siltstone, siltstone, mudstone, silty claystone, sandstone, limestone, and dolomite. In contrast, the cross-stratified facies is almost entirely sandstone or siltstone.

The Nugget Sandstone in Wyoming and northern Utah contains beds that were laid down in several environments–eolian, small lakes within dune areas, inland sabkhas, and streams (High and Picard, 1975; Picard, 1978). It had a hot and desert climate. It was trodden by dinosaurs, crocodiles, scorpions, and many other animals. There were small lakes, ephemeral with changes in the climate, within the dunes. Burrowers were locally abundant in what is now northeastern Utah, the Wind River Range, and probably elsewhere in Wyoming.

The building blocks and facings of some of the most attractive public buildings and homes in Wyoming and Utah have come from Nugget quarries. Nugget beds are short on fossils, other than burrows, but the warm colors–orange-pink, pale orange, grayish-orange, moderate reddish-orange, and moderate red–hold their color moderately well in these semi-arid to arid



Figure 9. Nugget Sandstone at Lander dome. Thinly bedded facies of Nugget unconformably overlies fine-grained facies of Popo Agie Formation. Light colored cliffs are Nugget cross-stratified sandstone.

climates. The cements–silica, the first formed, then calcite and dolomite–maintain the sandstone grains, though there is weathering of feldspars in the sandstone.

Gypsum Spring Formation

Though they have become caprocks over Nugget petroleum reservoirs in Wyoming and Utah, though evaporites overlie carbonates that contain one-half of the world's petroleum reserves, though gypsum is used to fashion wall board, which is made at two plants in the Bighorn Basin (Harris, 1993) and at a large operation in central Utah, Gypsum Spring rocks are little studied. It is a formation that droops and slumps and slides. The abundant gypsum beds (anhydrite at depth) in the lower part (**Figures 10** and **11**) lead to these instabilities and deformation within the formation. The anhydrite is interbedded with and underlies sealing claystone in Nugget oil-and-gas fields.

The contact between these earliest Middle Jurassic rocks (Bajocian) and underlying strata represents one of the most prominent unconformities in the Paleozoic and Mesozoic record of Montana, Wyoming, and Utah, said Downs (1949). Pipiringos and O'Sullivan, in their study (1978), assigned 2 to 3 million years to this unconformity ('J-1'), the contact between the Gypsum Spring Formation and the underlying Nugget Sandstone in Wyoming. They believed the lower unconformity (J-O) between the Nugget Sandstone and the underlying Popo Agie Formation or, in the Wasatch Mountains, the Nugget Sandstone and the Ankareh Formation, lasted much longer (7 to 10 my), though they assigned a Jurassic (?) and Triassic (?) age to the Nugget Sandstone. The duration of their J-0 cannot be well constrained between the slightly fossiliferous Popo Agie and Nugget Formations in Wyoming and Utah.



Figure 11. Gypsum Spring Formation at Lander dome, Wind River Basin. In the foreground, the photograph shows almost all of the formation.

From a maximum thickness of about 250 feet (75 m) in the northwestern corner of the Wind River Basin, the Gypsum Spring thins east and south and is absent in the southeastern part of the basin, largely because of truncation below the Sundance Formation (Burk, 1956). To the west, the Gypsum Spring thickens, gains limestone beds, and grades into the lower part of the Twin Creek Limestone.

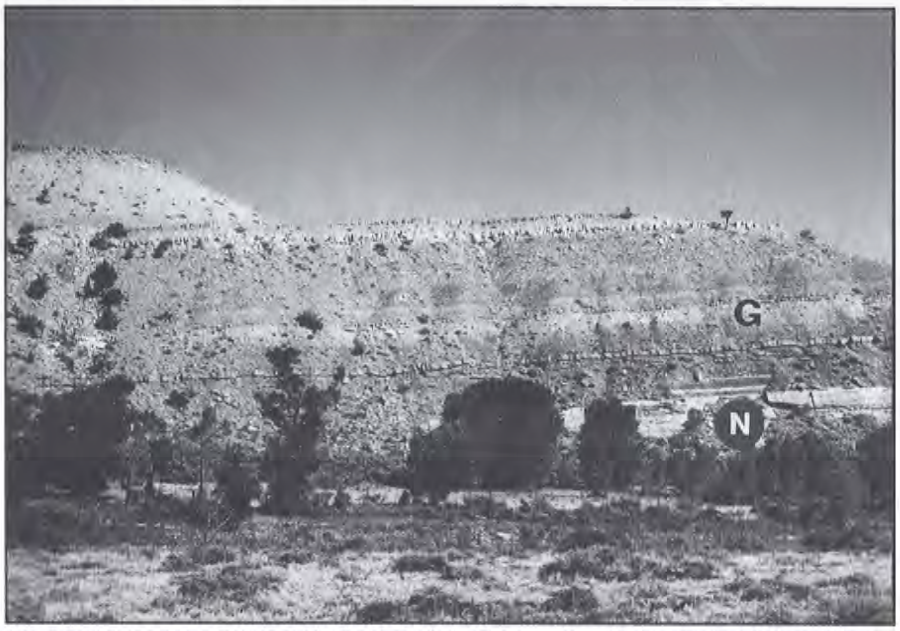


Figure 10. Uppermost Nugget Sandstone (N), Gypsum Spring Formation (G), and lower part of Sundance Formation (light-colored strata at upper left). Area in photograph is southeast of Dubois on U.S. Highway 287.

The lower part of the Gypsum Spring is the most widespread. The basal evaporite unit occurs at the wedge-edge in the Wind River Basin. Based on subsurface studies, Burk (1956) said that, "Throughout its extent in the Wind River Basin, the Gypsum Spring overlies the Nugget Sandstone." To the north, in the Bighorn Basin, the unconformity beneath the Gypsum Spring cuts successively down through Popo Agie, Jelm, Crow Mountain, Alcova, and upper Red Peak beds. About 35 miles (56 km) east of Sunlight Basin, in the northern Bighorn Basin, the Gypsum Spring lies on truncated Red Peak beds about 350 feet (105

m) below the top of the Red Peak Formation in complete Red Peak sections to the south. Loss of Gypsum Spring rocks by erosion occurs progressively on the east side of the Bighorn Basin, with the basal anhydrite unit present in all wells of Mills' (1956) cross section. Mills placed the Gypsum Spring-Red Peak contact at the base of the anhydrite in the subsurface.

In central Wyoming, the Gypsum Spring Formation generally is a lower unit of gypsum or anhydrite and red claystone and siltstone; a middle unit of limestone, dolomite, red and green claystone, and minor gypsum; and an upper fine-grained clastic unit that locally contains gypsum (**Figure 9**). Characteristic colors of the fine-grained "red beds" are moderate-orange pink, moderate red, grayish red, and pale reddish brown. Various shades of green are also characteristic of the fine-grained beds in the upper unit.

The marine limestone of the middle member preserves pelecypods and ammonites together with crinoids, corals, and gastropods (Love and others, 1945a). As in other parts of the world, an ammonite is very hard to find, perhaps requiring a day or much more of searching by a specialist. The limestone and some of the intertidal fine-grained clastic strata contain abundant foraminifera and ostracods (Peterson, 1957).

To the south, in northeastern Utah, and on the west in southeastern Idaho, the Twin Creek Limestone contains the Gypsum Spring Member at its base. This interval is partially equivalent to the Gypsum Spring Formation of the Wind River Basin.

On the pronounced unconformity (J-1) at the top of the widespread, wind-deposited, Nugget Sandstone sheets, the Gypsum Spring sea transgressed from the north into Wyoming and Utah. Thus began a period of marine evaporite deposition, probably on sabkhas, an Arabic word for salt flat. Thus continued, after a time gap of several million years, a period of arid and semiarid climates. Evaporation was greater than rainfall plus inflow of surface and subsurface water into the basin.

Sundance Formation

Before the invading Sundance seas, the stable Wyoming shelf was part of a regional low-lying erosion surface now preserved as the 'J-2' surface or 'J-2' unconformity. There was little relief throughout the whole region.

The lowermost few inches of the Canyon Springs Sandstone, the oldest member of the Sundance Formation, are characterized by weathered chert pebbles (Rautman, 1975). Pipiringos and O'Sullivan (1978) estimate that 80% of the chert is from the Gypsum Spring Formation. The Canyon Springs Sandstone in north-central Colorado is the youngest unit deposited over the J-2 surface. The oldest deposition on the J-2 surface occurred in the Williston Basin of Montana and Canada, and the western Dakotas about 161 Ma. The initial transgression of the Sundance sea to cover an area about 310 miles (499 km) east-west and 500 miles (805 km) north-south took approximately 5 my, which gives an average rate of transgression of about 60 to 100 miles (97 to 161 km) per million years (Pipiringos and O'Sullivan, 1978).

During Sundance time, sea dominated the land. The formation shows striking bands in the Black Hills of northeastern Wyoming and southwestern South Dakota where Darton first recognized their singularity (Darton 1899; Darton and O'Hara, 1909). The pleasing green and yellow Sundance outcrops, characterized by badlands of resistant sandstone and much thinner fine-grained intervals, are now easily recognized throughout much of central, south-central, and southeastern Wyoming (**Figure 12**).

Geologists' propensity–even their need–to recognize a lower part and an upper part of a formation is readily accommodated by the Sundance Formation. Each part is a major transgressional/regressional marine sequence (Brenner, 1983). The lower Sundance, from base upward–Canyon Springs Sandstone, Stockade Beaver Shale, Hulett Sandstone, and Lak Red Beds–is Callovian in age (Rautman, 1975). The upper Sundance–Pine Butte Sandstone, Redwater Shale, and Windy Hill Sandstone–is Oxfordian in age (Uhlir and others, 1988).

The advance of the lower Sundance sea is apparently represented by the Canyon Springs Sandstone and the lower part of the Stockade Beaver Shale, the prograding retreat by the upper Stockade Beaver Shale, the Hulett Sandstone, and the Lak Red Beds (Rautman, 1978). Rautman interpreted the transgressive phase to have been characterized by large, submarine sand waves (and/or tidal-current ridges), the regressive phase by nonmarine deposits on the southeast in western Nebraska and South Dakota, passing northwestward into lagoons, a barrier-island complex, a submergent bar complex, and into offshore silt and clay in central and northwestern Wyoming (see Rautman's figure 14, 1978).

The Stockade Beaver Shale grades vertically into Hulett Sandstone.

The Hulett Sandstone is inferred to have been deposited in the various suben-



Figure 12. Sundance Formation at Lander dome. All of the outcrop in the foreground is assigned to the Sundance. View is toward the northwest.

> *vironments of a barrier-island complex, including shoreface and beach, tidal channels, and back-barrier lagoon and tidal flats,*

said Rautman (1978, p. 2278). An ichnofauna of vertical, tubelike Monocraterion and Diplocraterion presumes very shallow water depths. Body fossils are rare. Wright (1973) reports only five bivalve genera from the Hulett.

Though an enigma, the Lak Red Beds are interpreted to be nonmarine, perhaps a windblown unit (Rautman, 1978). Like all of these early Mesozoic red bed facies, the Lak contains siltstone, sandy siltstone, claystone, and gypsum. It is unfossiliferous except for possible trace fossils. It lacks well-developed bedding and sedimentary structures. The wispy, mottled, swirled, sublenticular lamination resembles stratification and burrowed fabrics in the Red Peak Formation.

Brenner and co-workers suggested that sandstone and coquina facies of the upper Sundance are the deposits of offshore sand bodies (Brenner and Davies, 1973, 1974; Brenner and others, 1985). Uhlir and others (1988) said Brenner and his colleagues may be right in their interpretation of the upper Sundance in some parts of Wyoming, but are wrong for the Bighorn Basin where Uhlir and his colleagues believe the uppermost 50 to 165 feet (15 to 50 m) represent a tidal inlet, back-barrier shoal, and sandy tidal-flat that closes the marine Jurassic. As is true in many geological disputes, this one is courteous and unresolved, and seems to revolve around the question of which outcrops the protagonists have seen.

In the coquina facies, the shells are finely broken. Fragments typically are no larger than coarse sand. Rarely, whole disarticulated valves of *Camptonectes bellistriatus, Ostrea strigilecula,* and *Meleagrinella curta* (Imlay, 1956) rest in concave-down positions in the shell hash (Uhlir and others, 1988). The sandstone facies is fine, monocrystalline quartz sand "with 0-15% spheroidal glauconite and trace amounts of sand-sized dark-colored chert" (Uhlir and others, 1988, p. 739).

During the Late Jurassic, the stable Wyoming shelf had a paleolatitude between about 35° and 39°N. North was the equivalent of our northwest during the Jurassic. The climate then was generally warm and dry, as suggested by the abundance of red beds, evaporites, and shallow-water carbonate rocks (Brenner, 1983). Wind direction was west to east over the Western Interior. In the Oxfordian, the climate became more humid as moist westerly winds off the Pacific Ocean passed over newly uplifted elements on the west. The moist westerlies and moisture from the epicontinental sea combined to generate storms over the Western Interior sea (Brenner, 1983). These storms greatly influenced sediment dispersal patterns in the Sundance seas.

The transgressing and regressing shallow Sundance seas came in and went out over the Wyoming shelf, which was bounded on the west by the Utah-Idaho trough. On the east, the limiting feature was the Transcontinental Arch, a high area through the late Mesozoic (Rautman, 1978). Farther west (**Figure 13**), an active Andean-type subduction zone produced by the convergence of the Pacific and North American plates, raised positive areas along the western edge of the Sundance seas (Brenner, 1983).

The causes of fluctuating sea levels are diverse. Vail and others (1984) suggested that changes in ocean basin volumes may be the result of sea-floor spreading. Local and regional uplift also play a role in sea-level fluctuation. In the Late Jurassic, the east-west Belt Island trend, which originated in central Montana, was an effective barrier to invading marine waters from the north (Brenner and Davies, 1974). As the



Figure 13. Regional tectonic features and setting during the Late Jurassic (after Brenner, 1983).

western Cordillera rose during the Oxfordian, the sea was halted and reversed. Deposition exceeded subsidence, leading to a further regression.

Morrison Formation

Outcrops of the Morrison Formation at Como Bluff, southeastern Wyoming, made *Brontosaurus* a household word in the 1880s. The famous brontosaur (now apatosaur) quarry is in the middle part of the sequence in gray floodplain claystone, siltstone, and mudstone. It is about 100 feet (30 m) above the base of the Morrison.

Terrestrial environments represented in the Morrison Formation badlands (**Figure 14**) by alluvial fans, braided and meandering streams, and lakes succeeded the Sundance seas. Purple or green claystone and mudstone of the Morrison lie on fossiliferous limestone or sandstone or siltstone of the Sundance. Sundance sandstone may be highly glauconitic; green Morrison rocks owe their color to clay minerals. At some outcrops, the glauconitic rocks were reworked and glauconite occurs "several inches up into Morrison strata" (Moberly, 1960). Streams cut into the uppermost Sundance, scouring and channeling. Mud cracks and raindrop impressions on lower Morrison beds attest to the change to terrestrial conditions.



Figure 14. Badlands of Morrison Formation. View is northeast from Sheep Mountain, Bighorn Basin.

The Morrison lake basins were small compared with such large ancient basins as those occupied by Eocene Lake Uinta and Triassic Lake Popo Agie. The Morrison lakes were ephemeral. Rainfall was sporadic. The climate was strongly seasonal with high rainfall interrupted by brief dry seasons.

Subsurface studies of the Morrison Formation throughout Wyoming and northeastern Utah show that it is an upward-fining sequence. Outcrops of the Morrison characteristically display a lower unit of stream-deposited sandstone, siltstone and conglomerate, and an upper unit of more variable rocks, typified

by fine-grained strata–claystone, mudstone, and siltstone. There are lenses of channel sandstone and siltstone in the upper unit, and braided sandstone sheets of greater lateral extent. There also are carbonate beds and very limey siltstone. These formed in the small lake basins on the vast Morrison alluvial plain.

The fine-grained terrigenous rocks are variegated–red, maroon, green, gray, and brown. From a distance, the striking and characteristic color is purple–pale purple, grayish purple, and very dusky purple. Throughout the Rocky Mountain area, Morrison outcrops are more extensive than those of any other single formation, cropping out in badlands from New Mexico and Arizona on north into Canada. This vast low-lying area from north to south was about 1400 miles (2254 km) long, from east to west about 800 miles (1288 km) wide. Moberly (1960) has said that the land must have been near sea level because hypsometric curves "show very little of the earth's surface more than a few hundred meters in elevation." One wonders if, when Eldridge (1896) named the formation from outcrops near the village of Morrison in eastern Colorado, he had an inkling of the extent of the strata or how famous the beds would become.

Cretaceous Period

Introduction

I have looked at and mapped Cretaceous rocks in Wyoming and elsewhere in the Rockies for short periods of time throughout my professional career, beginning in Casper in the spring of 1950. My Cretaceous endeavors, however, have never paid well. My mistake, I now believe, came at the beginning, in working through part of the summer of 1949 as an unpaid assistant to a graduate student studying the Upper Cretaceous Frontier Formation in the Bighorn Basin. That study got me off to an unprofessional start from which I was never able to recover while working in Cretaceous terrains.

The Frontier Formation (**Figure 15**), consisting of sandstone, siltstone, shale, and coal, contains the deposits of several transgressions and regressions along the western margin of the Greenhorn Seaway. These sequences can be a nightmare to decipher because of internal unconformities, frequent facies changes, and deposition in every conceivable marginal-and shallow-marine environment (Siemers, 1975; Merewether and others, 1979; Merewether, 1983; Steidtmann, 1993). Further, the darn formation, though hospitable in many ways to humans, is also home to numerous rattlesnake families. My first Frontier mentor (Hunter, 1952) and I accommodated ourselves to the snakes, however, heeding their demands for space and ambling away from their dens to sit on potentially productive Frontier sandstone beds where we munched on raisin-and-carrot salad and ate the mayonnaise sandwiches my mother made for us to take to the Frontier hogbacks.

The literature on Cretaceous rocks of Wyoming is vast. In this review, the references cited represent only the tip of an iceberg. For the stratigraphic nomenclature, one can consult Love and others (1993), who have assembled an authoritative Phanerozoic chart with twenty-two Wyoming stratigraphic columns. Several papers, besides being excellent interpretations of aspects of the history, include extensive bibliographies on Cretaceous studies throughout the Rockies. These are: Dolson and others (1991); Gill and others (1970);



Figure 15. Giant "cannon-ball" concretions in Frontier Formation (maximum diameter is about 17 feet) at North Tisdale, north of Casper. Overlying Cody Shale in background.

Haun and Barlow (1962); Roberts and Kirschbaum (1995); Roehler (1990); Steidtmann (1993); and Weimer (1983). The sixteenth annual field conference guidebook published by the Wyoming Geological Association in 1961 treats Upper Cretaceous rocks. The *Geologic Map of Wyoming* (Love and Christiansen, 1985) includes a description of stratigraphic units and a long list of references.

Before turning to the Cretaceous history of Wyoming, I include a few comments about global Cretaceous events and life.

Chalk and the Cretaceous

Wrote H.G. Wells:

> *Who bears affection for this or that spadeful of mud in my garden? Who cares a throb of the heart for all the tons of chalk in Kent or all the lumps of limestone in Yorkshire? But men love England, which is made up of such things.*

In *The Time Machine*, Wells' Time Traveller may have been swept back into the past "into the abysses of the Cretaceous Sea."

What then of Cretaceous seas? "Cretaceous" comes from the Latin word *creta* for chalk (compare the French *craie* and the German *Kreide*). In 1822, J. J. d'Omalius d'Halloy, a long-forgotten Belgian, first applied the name to extensive chalk beds that form the white cliffs on both sides of the English Channel. Creteceous rocks occur on all the continents; worldwide transgressions flooded over older rocks on practically every land mass.

Most of the earth's chalk deposits–primarily made up of enormous numbers of ultra tiny plates consisting almost exclusively of the remains of floating (planktonic) marine microalgae, known as coccoliths–originated during the period. In Wyoming, chalk deposits are insignificant, and in most other regions where Cretaceous rocks occur, other sedimentary rocks greatly exceed chalk in volume. Pre-Cretaceous chalks do not exist.

Many years ago my older son and I , before giving ourselves up to the beauty of the Seine, the enigmatic allure of Mona Lisa, the pleasure of wandering through the Louvre, and the pull of Paris pastries, clambered over the Cretaceous chalk cliffs at Dover, climbing over strata laid down during the greatest submergence of the continents and in the most extensive epeiric sea the earth has known. Those onlapping Cretaceous seas, said Cloud (1988)–the products of expanding spreading ridges–inundated continents so leveled by weathering and erosion that detrital sediments were mainly fine-grained except where shed directly from existing or rising mountains.

Cretaceous life

Late Cretaceous vegetation bears little resemblance to that of the Triassic. Revolutionary changes occurred during the intervening Jurassic period. The changes between Late Cretaceous and present vegetation, however, are relatively minor.

In the Cretaceous, flowering plants (angiosperms), with their fruit-enclosed seeds, arose and came to dominate the land as they do today. Mammals and birds partly subsisted on the highly concentrated products of flowering plants–seeds, nuts, fruits–or on the insects that fed on these plants. Likely there was a connection between the rise of flowering plants and the later increase and expansion of mammals and birds.

Very recently, Archibald (1996) reported evidence for ungulate precursors that were present 85 Ma, pushing their ancestry back some 20 million years. These "zhelestids" (rat-sized ungulate forebears) were found in Asia in the Bissekty Formation by Lev Nessov, a Russian paleontologist. The finds suggest that ungulates may have first appeared there. If so, Asia was a cradle of mammal diversification, a place to look into mammalian origins. Near the Cretaceous-Tertiary boundary (65 Ma), noted Archibald (1996), "archaic ungulates appear in North America and shortly thereafter in South America, thus beginning a tremendous taxonomic and ecological radiation leading to the present ungulate diversity that ranges from antelope to elephants to whales."

Modern bony fishes arose in the Upper Jurassic, continuing their expansion through the Cretaceous. Modern types of clams and snails began to come into their own in the mid-and late-Mesozoic. Birds began their ascendancy, particularly toothed diving birds. From about 90 Ma, eleven kinds of diving birds are known, and sixteen kinds of water and shore birds, plus one land bird that occurs in later Cretaceous beds. One can see the culmination of archosauran evolution in the rich and wonderfully preserved Late Cretaceous dinosaur remains and in those remnants of other reptiles found in southwestern Canada (Alberta) and the northern United States, in Patagonia and Brazil, and in outer Mongolia (Cloud, 1988).

And then it was over. Following nearly 200 million years of varying success, and after recovering from the

Late Triassic extinctions, the remaining dinosaurs, pterosaurs, swimming reptiles, and all the archosaurs except for birds and crocodiles disappeared.

About 30% of invertebrate families became extinct at the end of the Cretaceous, among them ammonites, many kinds of bivalves, and planktic foraminifera. Ammonites, the trademark of Mesozoic marine life, declined in diversity for much of the Upper Cretaceous, noted Eldredge (1991). Somewhat similarly, Wiedmann and Kullman (1996) suggested that ammonoid extinction at the close of the Cretaceous was very similar to that at the Triassic-Jurassic boundary, "a long period of decline with a maximum number of extinctions long before the end of the Cretaceous." They contended that ammonoid extinction, therefore, was comparable to major extinctions at the other Mesozoic system boundaries, concluding that, "This extinction was unrelated to the K-T boundary event itself, which affected the oceanic surface plankton and was of either cosmic or volcanic origin." Comparing ammonoid diversity–based on the number of ammonite genera from 409 Ma to the present–with global sea-level changes, Wiedmann and Kullman (1996) showed a striking similarity, with extinctions related to times of regression and radiations occurring during transgressions. This is hardly a new idea (Moore, 1950; Newell, 1952) as Wiedmann and Kullman noted. Finally, Wiedmann and Kullman (1996) concluded that it is unnecessary to invoke a major cosmic impact to explain the final decline of ammonoids.

Ward's (1996) assessment of the final decline and extinction of the ammonoids contrasts greatly with that of Wiedmann and Kullman (1996); they could not be much further apart. Ward is a catastrophist, Wiedmann and Kullman, gradualists. Rather than a sterile period, Ward considered the end of the Cretaceous as one of the most evolutionarily fecund in the history of ammonoids. Following Collom (1986), and differing from those who interpret the record as a long period of ammonoid decline (Hallam, 1987; House, 1993), Ward (1996) suggested that even during the last of the Cretaceous (Maastrichtian) virtually no extinction took place, at least not until the very end of the age. He noted thirty-five ammonite genera in the lower Maastrichtian, twenty-seven in the upper. Following an episode of global anoxia in the seas at the end of the Campanian that did reduce ammonoid diversity, "ammonoid numbers stayed approximately constant until the end of the Maastrichtian" (Ward, 1996). Ammonites were then suddenly extinguished from the face of the earth.

Undoubtedly, said Ward, the final extinction of the ammonoids was brought about by the effects following the impact of the supposed Chicxulub comet or asteroid (Hildebrand and others, 1991) in the northern part of the Yucatán Peninsula in the Gulf of Mexico. Virtual extinction of plankton following the impact probably was the actual killing mechanism of the ammonites, he suggested. Ward also favored a collision event to account for the pattern of ammonoid disappearance at the Triassic-Jurassic boundary.

In contrast to Ward, Officer and Page (1996; also, see Meyerhoff and others, 1994) summarized the evidence and the argument against Chicxulub being an impact structure, contending instead that it is a volcanic sequence of the Upper Cretaceous. Said they, "It is not an impact melt sheet of Cretaceous-Tertiary age." However, in his deft and readable account of the great Cretaceous extinction–*T. Rex and the Craters of Doom*–Walter Alvarez marshals the evidence for the theory that Chicxulab–"the smoking gun"–is the impact crater of a comet or asteroid ending the Cretaceous with a series of terrible environmental disasters. Alvarez' book is the best account and summary of all the data favoring an impact that I have read. Certainly there was an impact event. There seems little doubt anymore of that. After seeing the collisions of fragments of comet Shoemaker-Levy 9 with Jupiter, I became, I must say, a true believer in the Cretaceous collision event 65 million years ago and in the likelihood that it doomed many groups of plants and animals–the Cretaceous mass extinction. Other causes may also have contributed to the K-T event but their role was less.

Large-scale environmental changes in Wyoming and the Rocky Mountain region lend support to the idea, further argued Officer and Page (1996), that a cosmic impact at the end of the Cretaceous is unnecessary to account for the massive extinctions. In brief, here is that scenario. During the late Maastrichtian, the relatively rapid regression of the epeiric sea caused profound changes in the paleogeography and foreland basin configuration. As the seaway retreated, nonmarine conditions prevailed throughout Wyoming and the Western Interior. Mountains arose and intermontane basins formed through Laramide-style deformation in the central United States. Internal drainage patterns were initiated. The lush swamps shrank and disappeared. The vast Western Interior seaway was gone, likely with a loss of almost all of the shallow-water regions and habitats. Western North America was the last haven for dinosaurs. At about the same time, environments suitable for shallow-water shellfish were restricted or nonexistent.

Intense volcanism and mountain making–the Laramide magmatic events from about 64 to 69 Ma–occurred in the Rocky Mountain region around the Cretaceous-Tertiary boundary (Snoke, 1993).

Extrusive activity was notable in the Ute, La Plata, and Elk Mountains and on the southwest in the Four Corners area, the common conjunction of Utah, Colorado, New Mexico, and Arizona. While worldwide sea level was falling, flood basalts named the Deccan Traps (located in western India) were deposited in a relatively short time period around the K-T boundary. The eruptions began during the magnetic polarity interval 30 N (normal) and continued during 29 N, but the bulk of the basalts were extruded during 29 R (reversed) (Officer and Page, 1996). It was during the 500,000-year-period of 29 R that the changes from Cretaceous to Tertiary microfossils took place.

More than a million cubic kilometers of Deccan lavas were ejected. Nothing like this cataclysmic extrusive upwelling has happened since. The intensive volcanism apparently was not limited to what is now India–as noted here–and was global in extent. Such volcanism ejected tremendous amounts of sulfur dioxide, carbon dioxide, chlorine and other gases into the atmosphere and stratosphere. McLean (1978) has suggested that a terminal Mesozoic greenhouse effect resulted from the carbon dioxide emissions. Keith (1980) postulated that the chlorine emissions destroyed the protective ozone layer, leading to an increase in ultraviolet radiation on the earth. The bare-skinned dinosaurs, he suggested, would have been at greater risk than birds and mammals, which are partially protected. Volcanism can not be dismissed as a possible contributing cause for catastrophic global events, but likely was not the principal cause of the K-T mass extinction. No reasonable link between impacts, volcanism, and mass extinctions is apparent. The seeming coincidence of them at the K-T boundary, however, cries for much further investigation.

Structural setting and seaway

Most of the present basin-and-mountain structure of the northern Rockies originated during the Laramide orogeny, beginning in the Late Cretaceous at about 70 Ma and ending in the Eocene at about 40 Ma (Dickinson, 1974; Snoke, 1993). Structural relief is extreme; the depth to Precambrian rocks in deep basins may exceed 25,000 feet (7500 m).

In the Early Cretaceous (Albian), the Wyoming shelf was cratonic, intermittently flooded during transgressions. During these first incursions, the sea laid down fine-grained sediment that became the Thermopolis-Skull Creek Shales.

The vast Western Interior seaway (**Figure 16**), connecting the Circum-Boreal sea with the proto-Gulf of Mexico during the Late Cretaceous, formed during a time of maximum eustatic sea level for the Phanerozoic (Vail and others, 1977; Haq and others, 1987). As previously noted, waters flooded the stable cratonic areas of the world. At its maximum extent, the Western Interior sea extended for 3000 miles (4830 km) from the North Slope of Alaska to northern Mexico and was about 1000 miles (1610 km) wide from central Utah to Minnesota (Kauffman, 1984). No good modern equivalent exists of such tremendous shelf seas for comparison.

The seaway lay within an asymmetric foreland basin bounded on the west by the Columbia-Sevier mountain belt and flanking foredeep, and on the east by the stable platform (Steidtmann, 1993; Roberts and Kirschbaum, 1995). As the foreland trough subsided, there was a eustatic (long term) rise in sea level, which Pitman (1978), Kauffman (1985), and Larson (1991) impute to a period of spreading-ridge and mantle-plume development. Greater spreading rates and the resulting formation of hotter, higher-standing oceanic crust displaced sea level a thousand feet or more above the present stand with flooding of ocean waters over low-lying parts of the continents.

The Cretaceous seaway began with separate north and south arms in early Neocomian time. With continued basin subsidence and sea-level rise, the two arms spread onto the foreland basin, connecting the proto-Gulf of Mexico with the Cretaceous circumpolar sea in middle late Albian time. Briefly, the arms disconnected, reconnecting during latest Albian-earliest Cenomanian. Rising and falling frequently, transgressing and regressing, the seaway remained on the continent for the next thirty-eight million years.

Because of the relatively flat sea floor, especially along its eastern shoreline, small fluctuations in sea level led to widespread, essentially synchronous strand-line migrations. The position of the western shoreline also varied widely; the interaction of varying relative sea level and the supply of sediment shed by the rising thrust belt on the west controlled its position (Steidtmann, 1993). In a paper by Gill and Cobban (1973), the western extent of marine beds–based on ammonite biostratigraphy–is used to determine the position of the western shoreline during the Late Cretaceous. By establishing correspondences between the ammonite zones of Obradovich and Cobban (1975) and North American Land Mammal "Ages," Lillegraven and Ostresh (1990) expanded this work, locating the marine-nonmarine transition along the western edge of the seaway as it developed during the Late Cretaceous.

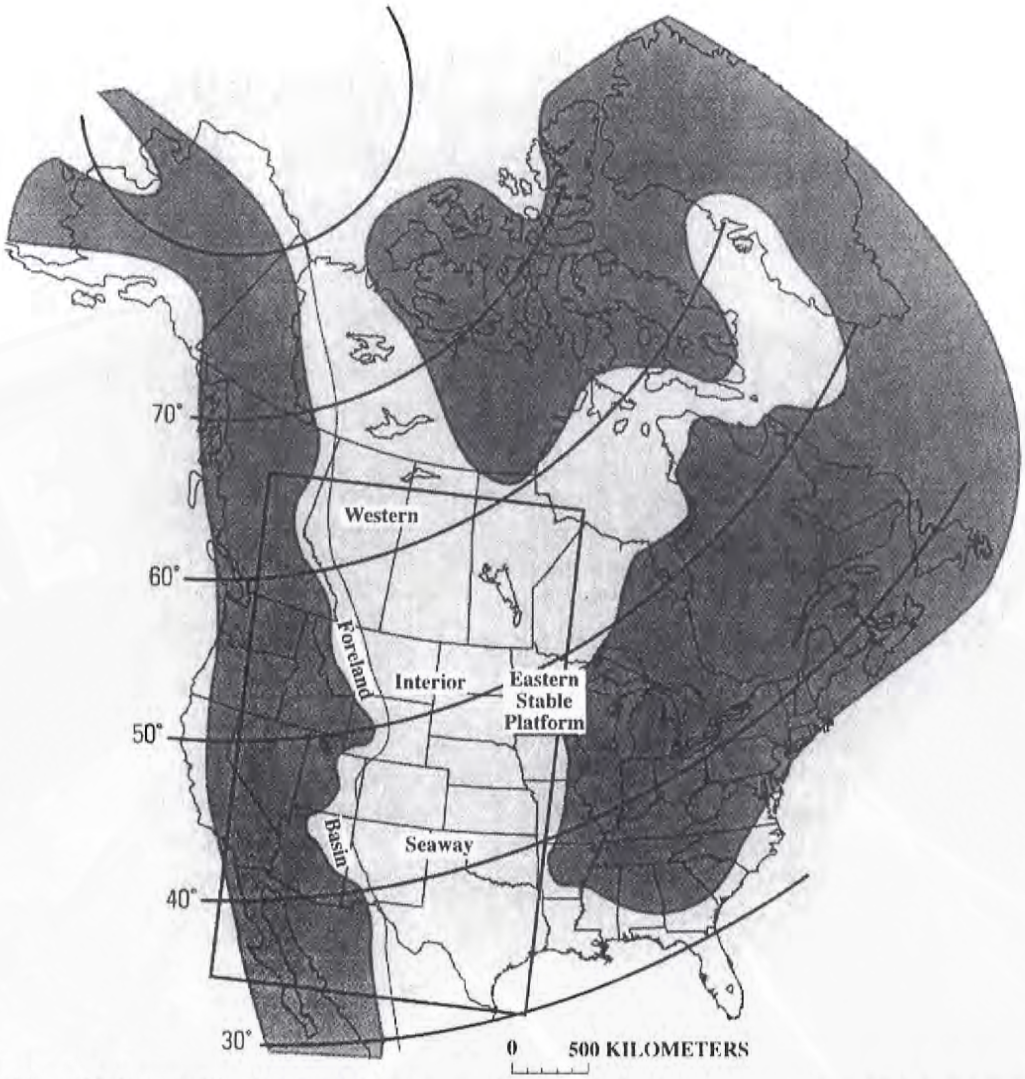


*Figure 16. Extent of the Western Interior seaway during peak late Turonian transgression (*Watinosceras *time). Note general locations of the Foreland Basin and the Eastern Stable Platform. The paleolatitudes are averages from about 90 Ma to 60 Ma. The figure is modified very slightly from a figure in Roberts and Kirschbaum (1995).*

Cyclicity

Trying to map anticlinal folds in the Mancos Shale–approximately equivalent in age to the Hilliard Shale of southwestern Wyoming–south of Shiprock, northwestern New Mexico, in the early 1950s, I learned the obvious: the repetitive Cretaceous strata record events of a cyclic sea. The Mancos in that part of the San Juan Basin dips at very low angles. I was tracing one-to two-inch thick orange siltstone beds–a rock type that later obsessed me when I began to study Triassic red beds. Because of the topography, it was necessary to skip frequently to different orange siltstone layers in younger or older cyclic sequences. I could not map one bed throughout the whole area, but converted each evening all the new measurements to a phantom horizon. The desert winds blew silt through the basin that spring much as winds probably blew silt over the Mancos sea some 88 to 84 Ma. In the field, I chewed grit every day south of Shiprock. By the end of a day, working with my data in the Shiprock Hotel Lounge toward midnight and beyond, I was ready to have an Orange Crush or two and I did so.

The Western Interior Seaway was cyclic. At a local scale–around Shiprock or in the Hillard Shale near Kemmerer, Wyoming–discriminating among effects of the many controlling variables is difficult, as Steidtmann noted (1993). Controls may have been long term (eustatic) or short term (climatic). Clastic cycles may also have been caused by autocyclic mechanisms–meandering of a channel in a river, tidal creek or submarine fan, subaerial floods, sub-aqueous

sediment gravity flows, channel switching on a subaerial or submarine fan or a delta (avulsion), storms and tidal ebb and flood (Miall, 1984).

Within the Cretaceous foreland basin, several orders of cyclicity have been recognized (Kauffman, 1985; Haq and others, 1987; Steidtmann, 1993). Two Cretaceous first-order cycles of 30 to 50 my (million years) duration are superposed in the overall record of flooding of the craton of the western United States from Jurassic to Paleocene time. Second- and third-order regressions removed part of the stratigraphic sequence, however, making it difficult to recognize first-order cycles. Second-order (6 to 10 my in duration) and third-order (1 to 5 my in duration) relative sea-level fluctuations produced landward and seaward movements of the strand line on the western and eastern basin margins. In the basin center, time-equivalent, large-scale cyclothems were laid down. These second- and third-order cycles are tied to major times of transgression and regression long recognized in the Cretaceous (**Figure 17**). Cretaceous sea-level curves have been published by Hancock (1975), Kauffman (1977, 1984, 1985) and Hancock and Kauffman (1979).

Ryer (1983) described fourth-order cycles, which are represented by smaller-scale progradational sequences on the western margin of the basin. Kauffman (1985) attributed these to short-term still stands or to a fall in relative sea level. Recently, Ryer (1983) suggested that clastic wedges on the western basin margin are of three general types: those associated with changes in relative sea level, those associated with varying depositional rates, and hybrids of the two end members.

Kauffman (1985) related even smaller-cycle cyclicity to changing climatic conditions. Three-to thirty-foot bedding cycles are common throughout the foreland basin, commonly expressed as claystone-limestone bedding couplets in the basin center and as prograding sandstone sequences truncated by black claystone beds near the basin margin. Based on radiometric dates, these couplets fall into three groups based on their supposed duration: 21 ka (thousand years), 42 ka, and 100 ka. Kauffman (1985) suggested that they correspond to Milankovitch cycles, reflecting orbital parameters of the earth. Barron and others (1985) presented sedimentologic, geochemical, and biological information indicating that claystone-carbonate cycles of these magnitudes reflect wet-and-dry climatic episodes.

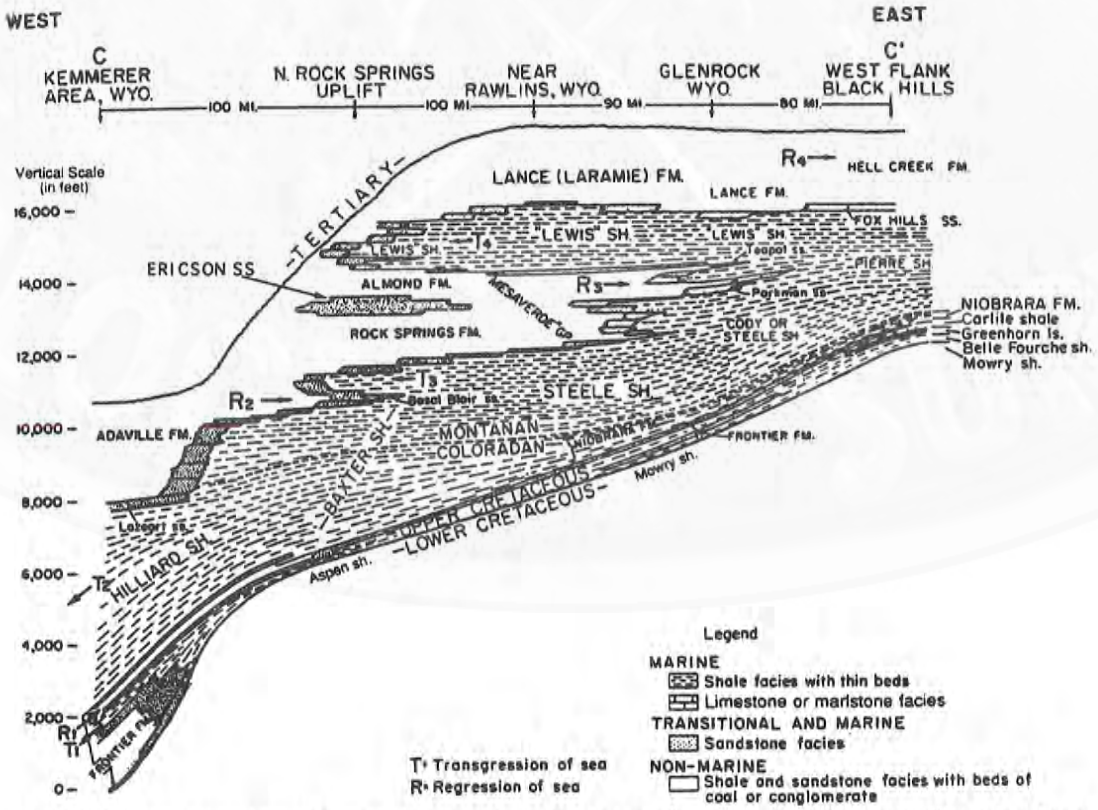


Figure 17. Restored section of Upper Cretaceous rocks from southwestern to northeastern Wyoming. From Weimer (1960).

Stratigraphic record

In the Early Cretaceous, sediment was derived from both sides of the foreland basin, but the greatest deposition was along the western margin. In the Late Cretaceous, most sediment likewise came from the west. Environmental changes from coastal plain to shoreline to marine shelf to deep marine controlled the lithofacies. The intertonguing of nonmarine strata on the west with marine strata in the basin's center is shown in **Figure 17**, a generalized cross section from the tectonically active western margin to the central basin area, northeastern Wyoming. Weimer (1983) discussed the influence of tectonics upon deposition and upon the related unconformities in the foreland basin fill, recognizing nine regional unconformities which he used to construct a restored section (**Figure 18**). The unconformities occur within nonmarine strata, within marine strata, and within sequences including both marine and nonmarine rocks. Where a large amount of time is missing, the approximate ages are those at the top of the subjacent unit (**Figure 18**). These ages are: (1) Late Neocomian to Early Aptian, 125 Ma; (2) Late Aptian to Early Albian, 112 Ma; (3) Albian, 101 Ma; (4) Early Cenomanian, 95 Ma; (5) Late Turonian, 90 Ma; (6) Late Turonian to Early Coniacian, 88 Ma; (7) Early Campanian, 80 Ma; (8) Late Campanian to Early Maastrichtian, 73 Ma; and (9) Late Maastrichtian, 66 Ma. Many of the unconformities can be correlated with sea-level changes. There is some uncertainty as to whether any particular unconformity is related to tectonism, sea-level change, or to both.

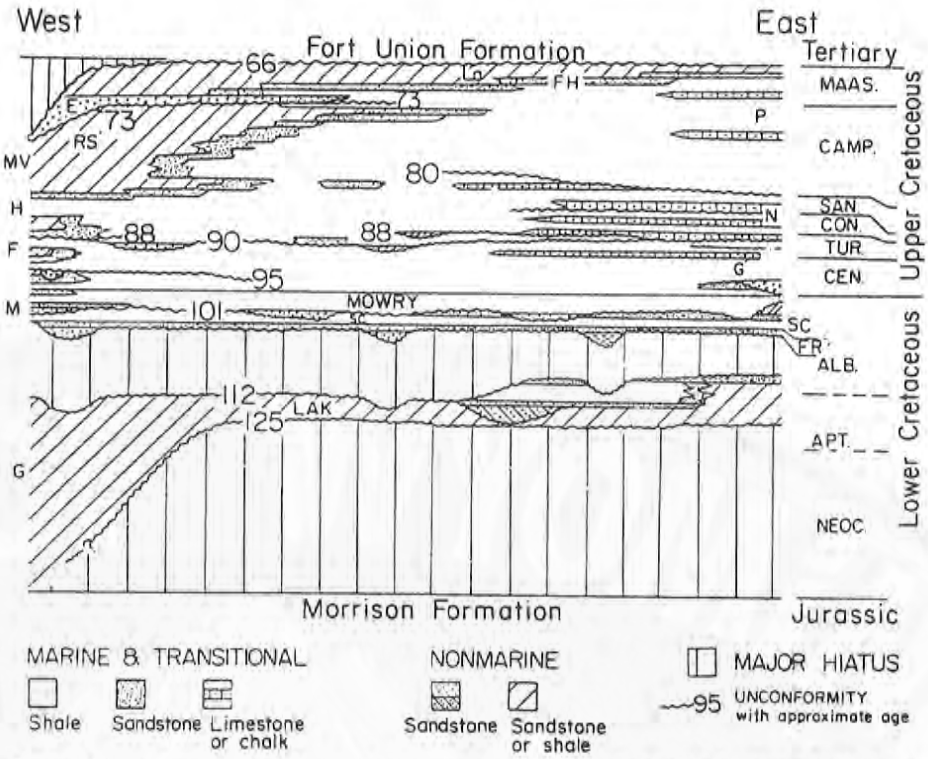


Figure 18. Diagrammatic restored cross section across the Cretaceous foreland basin from its western margin eastward to its geographic center. The stratigraphic positions of major intrabasin unconformities are shown by wavy lines that include approximate ages in millions of years. Strata are dominantly nonmarine in the western portion of the basin and dominantly marine in the geographic center of the basin. Unconformities are in three positions: those within nonmarine strata, those involving both marine and nonmarine strata, and those within marine strata. Wavy lines completely across the diagram represent times when the entire basin was subjected to subaerial or submarine erosion. Abbreviations for stages are: Neoc.=Neocomian; Apt.=Aptian; Alb.=Albian; Cen.=Cenomanian; Tur.=Turonian; Con.=Coniacian; San.=Santonian; Camp.=Campanian; Maas.=Maastrictian. Formations or groups to the west are: G=Gannett; M=Mowry; F=Frontier; H=Hilliard; MV=Mesaverde; RS=Rock Springs; E=Ericson; FH=Fox Hills; La=Lance. To the east, formations are: LAK=Lakota; FR=Fall River; SC=Skull Creek; G=Greenhorn; N=Niobrara; P=Pierre. The vertically ruled lines represent unconformities where a major hiatus is recognized. Ruling is omitted where the gap is small. Modified from Weimer (1983) to reflect new ages from J.D. Obradovich (in press).

Lower Cretaceous

During the Early Cretaceous the source of sediment was from both sides of the basin with the thickest strata accumulating along the western margin. In Wyoming, Neocomian through Early Albian sedimentary sequences range in thickness from 4000 feet (1200 m) on the west to less than 400 feet (120 m) over most of the state. Most of Neocomian time is unrepresented in Wyoming. On the west, however, the lower part of the Gannett Group–alluvial-fan and lake deposits that formed adjacent to the western highlands (Eyer, 1969; Furer, 1970)–may be as old as Late Neocomian. During the Early Aptian, gravels that characterize the Cloverly Formation spread eastward, laying down the first Cretaceous sedimentary record over much of the state. Middle and Late Albian beds are about 1500 feet (450 m) thick on the west and less than 300 feet (90 m) thick on the east. During this time, the first Cretaceous marine invasion–the transgressive phase

of the Kiowa-Skull Creek T-R (transgressive-regressive) cycle–advanced across Wyoming. The dark gray to black claystone of the Skull Creek Formation in northeastern Wyoming, and the Thermopolis Shale over most of the state, represent this depositional episode. The transgression was short-lived; a regional lowstand surface of subaerial exposure and erosion formed during the following major regression.

By 1990, the Lower Cretaceous Muddy Sandstone and equivalent strata had produced more than 1.5 billion barrels of oil-equivalent hydrocarbons (Dolson, and others, 1991). The Muddy, which is the most prolific stratigraphic-trap reservoir in the Rocky Mountains, may be the uppermost formation of the Lower Cretaceous in Wyoming (see Love and others, 1993). It, in turn, is overlain by the Mowry Shale and it is underlain by the dark gray-black Thermopolis Shale.

According to Dolson and others (1991), the Muddy Sandstone contains four facies. Facies one is fine-grained to conglomeratic lithic sandstone and quartzarenite with abundant fusain coaly debris, claystone rip-up clasts, erosional basal contacts, and trough- and planar-cross-stratification. Commonly, this facies represents point bar, braided stream, or tidal channel deposits.

Facies two is fine-to very fine-grained quartzarentite (in places the sandstone becomes arkosic) with gradational or sharp bases, moderate to abundant trace fossils, and horizontal to low-angle laminae. Locally, the sandstone is cross-stratified. Generally, such sequences represent beach, barrier island, or offshore-bar settings.

Interbedded, black, organic-rich, fissile claystone, coal, and waxy-green claystone (comparatively rare) interbedded with thin-bedded sandstone showing ripple- or wavy-stratification comprise facies three. Such sequences reflect deposition within lagoon, overbank, and splay settings or in abandoned channel fills.

The fourth facies, noted Dolson and co-authors (1991), consists of "rhythmically-bedded couplets of sandstone and fissile black shale," which they interpreted as a turbidite or density-flow deposits.

In their exhaustive study, they suggested that Muddy production is controlled primarily by unconformities formed during a relative sea-level lowstand, which is typical of many cratonic sequences elsewhere. The Muddy consists mainly of remnant ancient hills, valley networks, and onlapping marine units, all of which produce hydrocarbons. Reservoir rocks occur in ancient hills of older marine sandstone, younger valley fills and associated alluvial-plain channel sandstone, and in transgressive marine deposits. Seven of the ten paleodrainage basins they studied produce petroleum. The unproductive basins were insufficiently buried to form hydrocarbons or they were not located along migration paths, said Dolson and co-authors (1991).

On the south, the Mowry Sea was bounded by the Transcontinental Arch and was not connected to the proto-Gulf of Mexico. As the sea deepened, it became stagnant, leading to the formation of black claystone and petroleum source beds. The Aspen Shale, of western and northwestern Wyoming, and the Mowry Shale throughout the rest of the state, accumulated in this sea. The reader is referred to the paper by Reeside and Cobban (1960), in which they report on studies of the formation in the western United States and Canada. In Wyoming, the conspicuous, siliceous, silver-gray ridges of the Mowry contain porcellanitic shale, porcellanite, and bentonite, all products of the long-term volcanism. Mowry strata have yielded brackish-water and marine invertebrates; the formation is noted for disarticulated fish remains. Fish scales are common.

As studies continue, the rock package assigned to the Lower Cretaceous in Wyoming is becoming thinner and thinner. In the early 1980s, Merewether (1983) and Weimer (1983) and most other workers assigned all of the Mowry Shale to Late Albian. Based on a reassessment of ammonite zones and their ages (Cobban and Kennedy, 1989), Love and others (1993) suggested that the Mowry is at least partly Early Cenomanian, placing that part in the Upper Cretaceous.

Upper Cretaceous

Upper Cretaceous strata are 18,000 feet (5400 m) thick in the western foredeep and 16,000 feet (4800 m) thick in south-central Wyoming, diminishing to less than 4000 feet (1200 m) in the central-western part of the state. During the transgressions and regressions of the Late Cretaceous, a north-south belt of intricately intercalated marine, deltaic, barrier-island, and coastal-plain sediments accumulated across Wyoming, separating dominantly marine beds on the east from dominantly nonmarine and marginal-marine beds on the west.

In the Mowry seaway, eustatic sea level continued to rise, flooding over the Transcontinental Arch in Early Cenomanian, reestablishing a through-flowing connection with the proto-Gulf of Mexico, and forming the Greenhorn Sea (Kauffman, 1984, 1985; Weimer, 1983). During early Turonian, the Greenhorn seaway reached the highest eustatic sea level of the Creta-

ceous; clastic sediment that became the Frontier Formation spread eastward over the Aspen and Mowry Shales along the western seaway margin. Merewether (1983) suggested that Frontier deposition began as early as latest Albian in the western part of the state, continuing into the Coniacian. Weimer (1983) contended, however, that all of the Frontier Formation lies above the Albian-Cenomanian boundary. In eastern Wyoming, formations equivalent in age to the Frontier include the Belle Fourche Shale, Greenhorn Formation, and Carlile Shale (Merewether and others, 1979).

Because of internal unconformities and facies changes related to transgressions and regressions along the western margin of the so-called Greenhorn Seaway, the Frontier is a greatly variable formation in rock types, age, and thickness. Merewether (1983) presented evidence that Frontier transgressions and regressions were caused by local tectonics as well as by fluctuations of sea level. The areal distribution of some nearshore marine and nonmarine beds and the spatial and time magnitude of hiatuses indicate that there were tectonic high areas at 90 to 91 Ma in western Wyoming, 89.8 Ma in southeastern and northwestern Wyoming, and 89.3 Ma in southeastern, central, and northwestern Wyoming (Merewether and Cobban, 1986).

In the eastern Uinta Mountains, the Frontier Formation is only 100 to 200 feet thick (30 to 60 m), but just west of the Uintas at Coalville, Utah, it is more than 4400 feet (1320 m) thick and contains conglomerate as well as sandstone, claystone, and coal (Hansen, 1975). Farther west, near Salt Lake City, the Frontier is more than 8700 feet (2610 m) thick. Throughout most of the Uinta Mountains the Frontier is twofold in character: a slope-forming clayey unit below, and a cliff-forming sandstone above (**Figure 19**), which is mostly some shade of gray or brown. Like the Dakota Sandstone in the Uintas, the Frontier consists of beach-laid sandstones, near-shore marine sandstones, and coastal-swamp beds. Oysters and other oyster-like shellfish are characteristic of the Frontier, but there are other mollusks as well, particularly in the lower part of the formation. At Coalville and at Kemmerer, Wyoming, the Frontier contains large coal reserves.

As the Frontier Formation accumulated in northeastern Utah, newly formed mountains that shed sediment toward the east bordered the foreland basin. Apparently the borderland in Utah was not far west of the present Wasatch Range. Thinning eastward, the Frontier gives way and intertongues with deep-water fine-grained strata–the Mancos, Hilliard, and Baxter Shales.

As the Niobrara T-R cycle began (**Figure 17**), marine conditions prevailed throughout most of Wyoming. During late Turonian and early Coniacian, the Niobrara transgression spread quickly, flooding the seaway with the second highest sea-level highstand of the Cretaceous (Steidtmann, 1993). Unlike other Cretaceous transgressions, this one was very prolonged, relieved by moderate third-order sea-level fluctuations. In the west, the Hilliard and Baxter Shales were deposited; in central Wyoming, the Cody Shale. In eastern Wyoming, the chalks and marlstones of the Niobrara Formation were laid down during this time, probably recording periods of transgression and peak flooding of the Niobrara sea. In the western foredeep, the maximum thickness of this sequence is about 6000 feet (1800 m). It thins to approximately 800 feet (240 m) in the east. Compared in volume with Cretaceous chalks found elsewhere in the world, the Niobrara chalky intervals are minuscule.



Figure 19. Lower Cretaceous sequence exposed on southeastern flank of the Uinta Mountains, Dinosaur National Monument, Utah. Morrison (Jurassic) Formation (right center), then successively the Cedar Mountain, Dakota, Mowry, and Frontier Formations.

While fine-grained sediments continued to accumulate in the east, regression of the Niobrara Sea is marked by the early Campanian appearance of clastic sediments which became the Mesaverde Group over much of the western and central part of Wyoming. Steidtmann (1993) noted that the position of the boundary separating the rocks of the Niobrara and Claggett T-R cycles is unclear, but it probably occurs at the base of the Rock Springs and Haystack Mountains Formations (**Figure 20**) in the lower part of the Mesaverde Group. Along the Wyoming-Utah state line, the Mesaverde is almost 7000 feet (2100 m) thick, thinning eastward, and pinching out into the Cody and Steele claystones in eastern Wyoming. The stratigraphic relationships created by changes in basin subsidence, sediment supply, eustasy, and change in position of environments are complex; they have led to a complicated and frequently changing stratigraphic nomenclature (Roehler, 1990; Steidtmann, 1993; **Figure 20**).

In southwestern Wyoming, a regional unconformity separates the lower part of the Blair and Rock Springs Formations, and the lower part of the Ericson Sandstone, from the upper Mesaverde Group, which includes the upper part of the Ericson Sandstone and the Almond Formation (**Figure 20**). The lower Mesaverde progrades to the east, ending with the distal Allen Ridge Formation, which overlies the Haystack Mountains Formation in south-central Wyoming. The upper Ericson Sandstone and the Pine Ridge Sandstone (and the equivalent Teapot Sandstone to the north) were laid down on the erosional surface, which was tectonically elevated and in places truncates as much as 1700 feet (510 m) of the lower Mesaverde (Roehler, 1990). At the top of the Mesaverde, the Almond Formation intertongues eastward with the Lewis Shale (**Figure 20**).

Depositional settings recognized in the Mesaverde are many: stream, coastal-plain, tidal, marginal marine, near-shore. There were extensive fluctuations in sea level and in consequent erosion during Mesaverde deposition and throughout the Cretaceous foreland basin at other times.

Bentonite, a clay deposit formed by the diagenetic alteration of volcanic ash, is abundant within the Upper Cretaceous (Harris, 1993), but most especially so in the Mowry and Frontier Formations. In Wyoming, the Ardmore bentonite (early Campanian) is a bed or series of beds about 5 feet (1.5 m) thick interbedded with strata of the Mesaverde Group and Cody Shale. It is a time-stratigraphic marker over much of the northern Rocky Mountain region, reflecting large-scale volcanism in west-central Montana approximately 80 Ma.

Because of its physical properties–high viscosity, cation exchange capacity, expansion when wet–bentonite has many applications. It is useful as a binder in pelletizing taconite (siliceous iron ore); as a lubricant, sealant, and thickener in drilling fluid; as a sealant in waste-isolation ponds; as a coagulant and cation exchanger in water and waste treatment; and as a mineral filler in many products (Harris, 1993). Bentonite is also being used to sequester mycotoxins in livestock feed grains. Seemingly this is a growing market because of the growing problem with mycotoxins. Before 1991, the greatest use of bentonite was in drilling fluid. Now, most bentonite is used in taconite pelletizing.

It is of interest to builders and homeowners that decorative stone has been quarried from the Mesaverde south of Rawlins. It was used in a number of Wyoming buildings, including the State Capitol, and in the Union Pacific Railroad Depot, Ogden, Utah (Harris, 1993).

The Lewis Shale (**Figure 20**) in Wyoming is much younger than the Lewis Shale at the type locality in Colorado. It correlates with the Bearpaw Shale of Montana (Weimer, 1960). In south-central Wyoming (Gill and others, 1970), the Lewis attains a maximum thickness of 2600 feet (780 m), and is characterized by dark-gray shale, light-colored sandstone, and coal. Along northeast-southwest lines of section the Lewis shows a pattern of submarine topography that Asquith (1970) interpreted as shelf, slope, and basin marine. Winn and others (1985) recognized shelf, delta-front, and turbidite deposits in the Lewis. Perman (1990) also recognized deltaic deposits in the Lewis, interpreting the Dad Sandstone member as a shelf sand-ridge.

The Bearpaw regression represents the last retreat of the epicontinental seaway from the Rocky Mountain region. Deposition during this interval records the transition from marine to nonmarine settings. The Fox Hills Sandstone, which intertongues with and overlies the Lewis Shale, represents this time interval in Wyoming. Except in south-central Wyoming where Fox Hills strata represent delta deposits, the Fox Hills Sandstone was laid down on a sandy shoreline as the sea retreated to the northeast. While the sea regressed eastward, nonmarine conditions were dominant, and the sediment that became the Lance Formation–gray sandstone, gray-and-brown shale, coal–and equivalent units was deposited (Steidtmann, 1993). In southeastern Wyoming–at Rock River, for example–equivalent beds are named the Medicine Bow Formation. Conglomerate of the Harebell Formation and part of the Evanston Formation were deposited in the far west. In central Wyoming, the Lance is up to 5000 feet (1500 m) thick (Keefer, 1965; Gillespie and Fox,

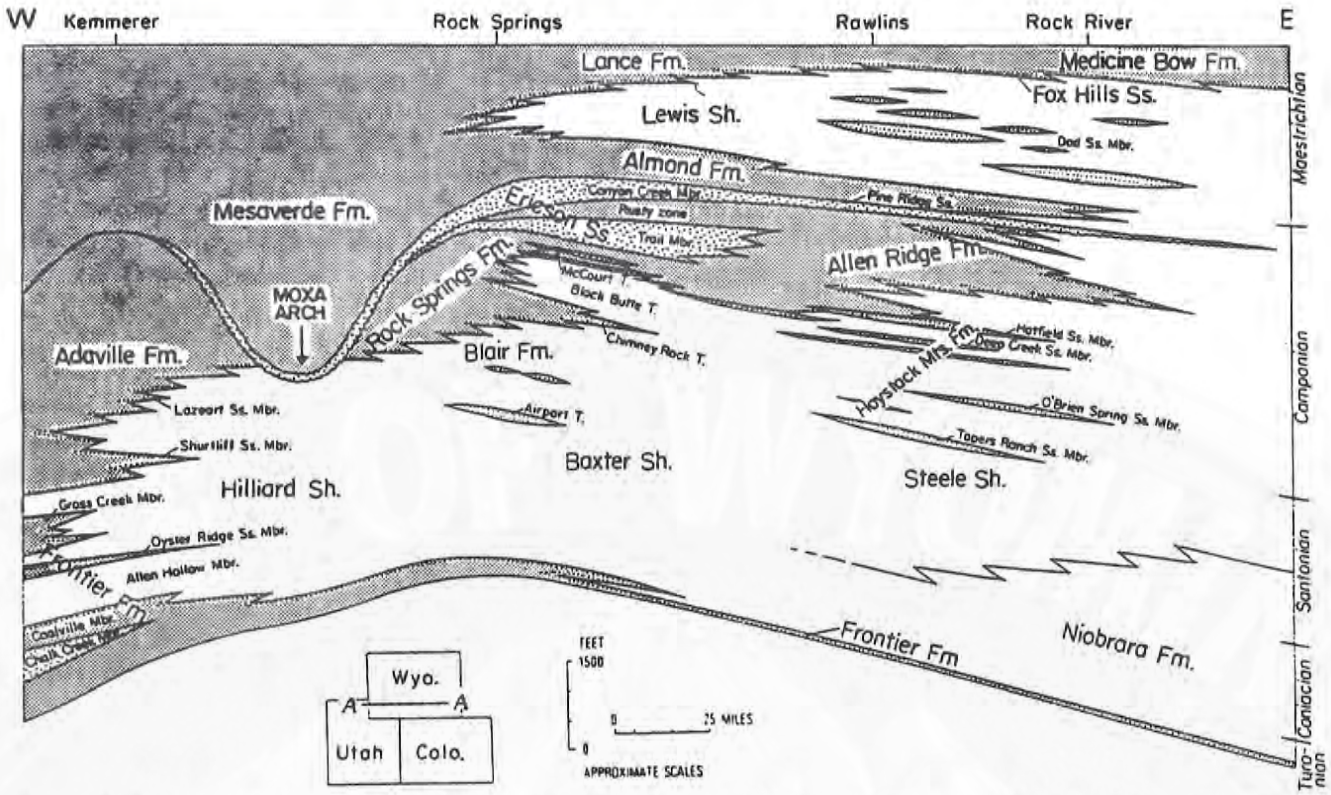


Figure 20. Cross section of restored Upper Cretaceous strata across northern Utah and southern Wyoming. Rocks of nonmarine origin are shaded; alluvial-plain and marine (or shoreline, shelf, slope, and basin) sandstone and siltstone units are shown by dot pattern; marine shale and limestone are unpatterned. From Roehler (1990).

1991), recording the transition from coastal-plain and delta-plain settings to stream deposition as the sea withdrew (Flemings and Nelson, 1991). The Lance equivalent Medicine Bow Formation is 6300 feet (1900 m) thick in southeastern Wyoming. Many geologists have cited thickness variations and unconformities in the Lance as evidence for the influence of Laramide tectonism on depositional histories (Keefer, 1965; MacLeod, 1981; Flemings and Nelson, 1991; Gillespie and Fox, 1991). Lillegraven and Ostresh (1988) demonstrated the influence of early Laramide uplifts on drainage patterns during Lance deposition. Dickinson and others (1988) cited the presence of local depositional centers and source areas as indicators of Laramide folding and faulting during the Maastrichtian. On the west, conglomerate intervals originated in and adjacent to the thrust belt. Gravels were shed in the west by interior, ramp-supported uplifts during the Maastrichtian; these are represented by conglomerates in the Evanston and Harebell Formations (Salat and Steidtmann, 1991).

Summary

The Triassic and Jurassic endured for 101 my, or 115 my, or perhaps something in between, depending on which authorities one consults. The maximum thickness of early Mesozoic rocks, using the thickest sequence for each formation, reaches about 4000 feet (1200 m) for Wyoming and northeastern Utah. Using 8000 feet (2400 m) of sediment as a maximum thickness before compaction and lithification, and about 100 my for its deposition, the depositional rate would have been 0.00008 feet per year. The 12,500 years necessary to accumulate a foot of early Mesozoic rock nearly exceeds man's habitation of North America.

In the Triassic and Jurassic, roughly equal amounts of sediment were laid down in continental settings–lake, stream, eolian–and in shallow marine or deltaic-plain settings–delta, beach, marsh, tidal flat, shallow shelf. The Nugget and Navajo Sandstones show incomparable outcrops and details of one of the world's great ancient eolian sandstone sheets. No geologist traveling through the Western United States could have a better opportunity to see a record of dominantly eolian deposition. Similarly, though much more difficult to study, the Popo Agie lake beds offer one of the world's most interesting fine-grained formations.

Some of the earliest interpretations of tidal flat deposits were made on the Red Peak and Crow Mountain Formations. The variety of primary sedimentary structures and bedding types in the Red Peak Formation, and in correlative beds in Utah, is enormous. The fine detail of lamellae, the contrasts between reddish brown and dark red and light-grayish green claystone, siltstone and very-fine sandstone, and the vivid expression of small-scale sedimentary structures has attracted students to these outcrops for generations. We should not forget the importance of the Morrison Formation in the geology of the West. Dinosaurs and uranium deposits, the largest ancient beasts and the most devastating explosions in our time–what a juxtaposition they are in Morrison channel conglomerate and sandstone.

Clastic rocks are overwhelmingly dominant in the early Mesozoic stratigraphic columns of the region. In order of decreasing abundance, they are: sandstone, siltstone, claystone, carbonates, evaporites, and claystone-pebble conglomerate. Siltstone is nearly as common as sandstone.

During the early Mesozoic, the Wyoming shelf was emergent or covered by a very shallow sea. The continentality of the Triassic and Jurassic sequence sets it off from preceding and succeeding geologic periods. Major petroleum source beds occur in the Permian (Retort and Meade Peak Shale Members of the Phosphoria Formation) and in the Cretaceous (Thermopolis, Mowry, Cody, Aspen, Skull Creek, and Carlile Shales; Frontier and Niobrara Formations) in the Rocky Mountains; they are absent or very minor in the Triassic and Jurassic. Permian and Cretaceous seas were deeper, covered far more of the craton, and persisted longer; marine organisms were, of course, much more abundant. Whereas early Mesozoic environments were dominantly oxidizing, Permian and Cretaceous environments were dominantly reducing.

Red beds (including varicolored badlands strata) constitute about 55% of the Triassic and Jurassic stratigraphic sections. The Red Peak, Crow Mountain, and Gypsum Spring red beds–approximately one-half of all the red beds in the sequence–accumulated primarily in shallow marine environments. The lateral persistence of facies has struck nearly all students of Triassic and Jurassic history in the Rocky Mountain region. Many such instances are noted here.

The Cretaceous continued for about 80 my, almost 45% of Mesozoic time. Cretaceous rocks are 23,500 feet (7050 m) thick in their thickest sequences in western Wyoming, 5500 feet (1650 m) of Lower Cretaceous, and 18,000 feet (5400 m) of Upper Cretaceous. Using 47,000 feet of sediment before compaction, the average depositional rate was about 0.0006 feet/year. The approximately 1700 years required to add a foot of Cretaceous rock was about 7.5 times as quick as the depositional rate during the combined Triassic and Jurassic in Wyoming.

Major oil-and-gas reservoirs occur in the Lower Cretaceous Cloverly Formation and Muddy Sandstone and in the Upper Cretaceous Frontier Formation, the Shannon Sandstone Member of the Mesaverde Formation, and the Almond Formation (uppermost formation of the Mesaverde Group). The complex intertonguing of Cretaceous marine sandstone and claystone, the thick laterally extensive coastal-plain sequences, and the many associated source beds provide the conditions for large hydrocarbon accumulations. More than one-half of such production has come from reservoirs (mainly sandstones) in anticlines (De Bruin, 1993). Many of the traps, however, have a stratigraphic component that is related to facies changes, diagenesis, or erosional truncation, and was a significant factor in the origin of particular reservoirs (see Dolson and others, 1991). Particularly large hydrocarbon reservoirs occur in valley-fill Muddy Sandstone deposits, shelf-sandstone ridges of the Shannon and Sussex sandstones, Frontier Formation shoreface sandstones, and Almond Formation barrier-island deposits. Of great importance to Wyoming's future are the enormous reserves in low permeability ("tight") gas sandstones of the Mesaverde Group, western Wyoming.

Excepting Alaska, Wyoming has the largest in-place coal resources of any state (1.44 trillion tons), noted Moore and Shearer (1993). All of the Wyoming coal fields contain Cretaceous-age coal beds. The most widespread deposits are present in the Mesaverde Group–Powder River, Green River, Bighorn, Wind River, Jackson Hole, and Rock Creek coal fields–and in its western equivalent, the Adaville Formation (the Hams Fork coal field). Paleocene coals crop out in all the coal fields except for Black Hills and Goshen Hole. They are present mainly in the Fort Union, Ferris, and Hanna Formations (Moore and Shearer, 1993). Eocene coals occur in the Powder River coal field (Wasatch Formation) and in the Green River and Hanna fields. Cretaceous coals, the result of their formation in deltaic, coastal plain, and other nearshore settings, are generally laterally extensive, but they are thin (less than 9 feet thick) and relatively high in sulfur (frequently greater than 1%) because these layers were succeeded by marine deposits.

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This article is dedicated to my friend Lee R. High, Jr., field mate and colleague for eight wonderful seasons. Lee and I co-authored the memoir *Sedimentary Structures of Ephemeral Streams*, twenty-six research articles, and sixteen abstracts, many of them on early Mesozoic history, stratigraphy, and sedimentary petrology in the Rocky Mountains.

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"The Gangplank" west of Cheyenne. View is eastward toward Cheyenne and the High Plains showing Interstate Highway 80 (to right) and the Union Pacific Railroad (to left). The aggregate quarry visited on the field trip to southeastern Wyoming is located in the bottom right of photograph. (Original color photograph by R. D. Miller, U.S. Geological Survey, from U.S. Geological Survey Bulletin 1493; reprinted with permission of the Theodore Roosevelt Nature and History Association.)

Glaciation of the Wyoming mountains and Pleistocene permafrost in the basin floors

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Abstract

Today small glaciers are restricted to Wyoming's highest peaks and intense frost action is limited to alpine sites. During the Pleistocene glacial maxima, ice flowing from cirques carved glacial troughs in the mountains and deposited tills and outwash well out on the margins of the adjacent plains during five major advances, named from type localities in the Wind River Range. Knob-and-basin topography is prominent on Wisconsin Pinedale moraines, more subdued on the Illinoian Bull Lake moraines, and largely absent on the progressively older tills.

Evidence of former tundra on the basin floors is provided by sand and pebble-filled casts of former ice-wedge polygons and relics and sand-wedge polygons. Their sites are mainly exposed where natural deflation and excavations along highways and pipeline trenches have removed post-Pleistocene covers of alluvial and eolian deposits. Other origins are rejected because: 1) these structures cut Pleistocene gravels devoid of swelling-and-contracting clays, as well as a variety of bedrock types which include shales, sandstones, and redbeds; and 2) similar host materials in the lower and warmer plains to the south of Wyoming lack such polygonal forms, although they are present at the 8000-foot levels of North Park in Colorado. During the glacial maxima, mean annual temperatures on the Wyoming basin floors (some 4500 to 8500 feet above sea level) are estimated as 10° to 16°C colder than the present–based on the temperatures where ice and sand wedges are now forming in Alaska.

Introduction

The following discussion is a synopsis of the many glacial studies in Wyoming, and the evidence for Pleistocene permafrost in the high-floored intermontane basins and eastern plains.

The Pleistocene climatic history of Wyoming is recorded by prominent glacial landforms on and near the mountains (which form 29% of the state's 97,281 square miles), and by widespread tundra relics exposed where covers of Holocene deposits have been eroded off the basin floors (which occupy 71% of the state) (**Figure 1** and **Table 1**).

The only present-day glaciers in Wyoming are relics of neoglaciation, a "Little Ice Age," from about 350 to 150 years ago when glaciers expanded on the state's highest summits. Since then the many small glaciers (**Figure 2**) on the peaks–as much as 13,804 feet high in the Wind River Range and 13,772 feet high in the Teton Range–have markedly receded; have largely disappeared on the 13,167-foot-high Bighorn Mountains; and are long gone from the 12,013-foot-high Snowy Range in the Medicine Bow Mountains. However intense frost action in the alpine zones–marked by shattered blocks, solifluction, and other periglacial phenomena–is an ongoing phenomenon above the 10,500-foot upper limits of the forested mountain flanks. The mountains descend to the sage-grasslands on the state's 5000- to 7500-foot-high intermontane basins and eastern plains.

In the Pleistocene epoch, or "Great Ice Age" from 1.65 million to 10,000 years ago, Wyoming had five or more glacial episodes with intervening warm intervals. At the glacial maxima, ice flowing from cirques in the high peaks eroded pre-existing mountain valleys into U-shaped canyons (**Figure 3**) and deposited end mo-

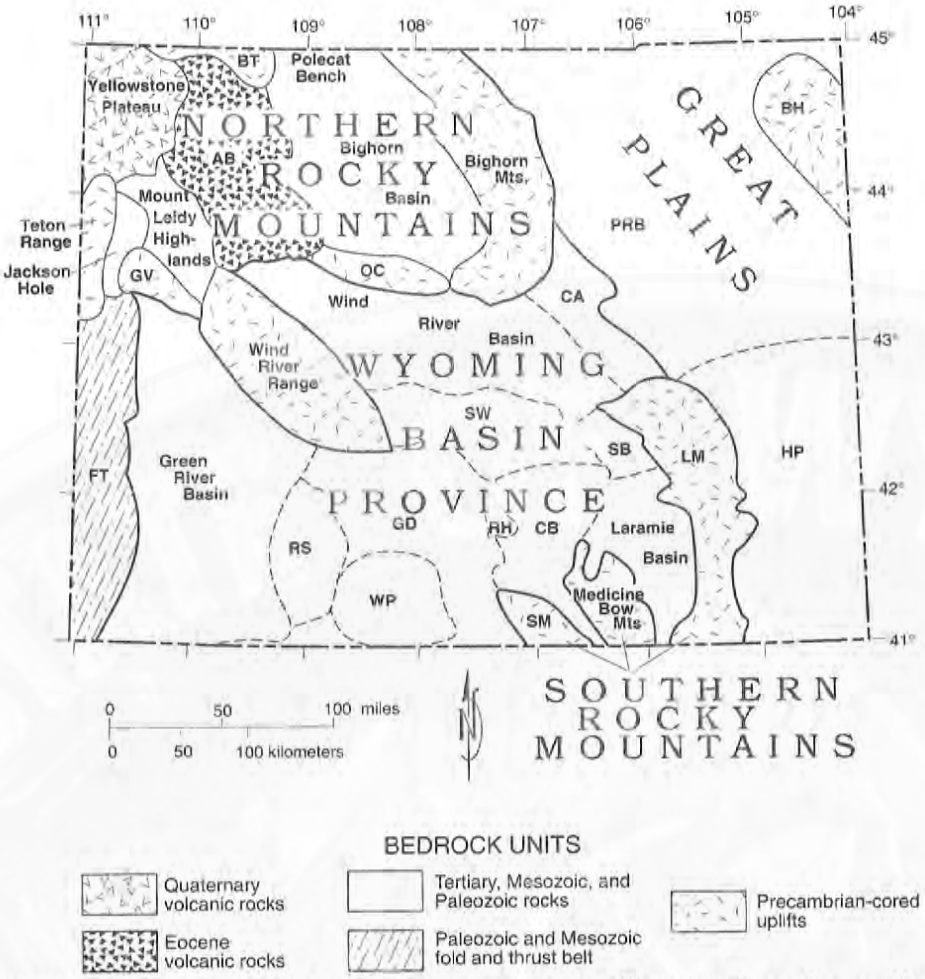


Figure 1. Geomorphic provinces, mountains, and basins of Wyoming. Those discussed in the text are named on the map. The others, designated by abbreviations, are named in ***Table 1****.*

raines and outwash plains beyond the mountain fronts (Mears, 1974). During the glacial episodes, there is good evidence that the intermontane basin floors and the Great Plains of Wyoming were middle-latitude tundras–shrub-grasslands underlain by permanently frozen ground of the sort now found in Alaska, northern Canada, and Siberia.

The glacial sequence

In his classic study of the post-Cretaceous origin and evolution of the mountains in central and northwestern Wyoming, Eliot Blackwelder (1915) proposed three major glacial episodes–Buffalo, Bull Lake, and Pinedale. His Buffalo glaciation, named from a locality near the headwaters of Buffalo Creek (which flows from the Mount Leidy Highlands into Jackson Hole, **Figure 1**), was represented by widespread relics of bouldery deposits overlying the sedimentary rocks

Table 1. Key to Wyoming glacial and ice wedge sites. See ***Figure 1*** *for locations of features.*

MOUNTAINS

Glaciated		Unglaciated
Absaroka Mountains (AB)	Mount Leidy Highlands	Black Hills (BH)
Beartooth Mountains (BT)	Owl Creek Mountains (OC)	Laramie Mountains (LM)
Bighorn Mountains	Sierra Madre (SM)	Rawlins Hills (RH)
Fold-Thrust belt (FT)	Teton Range	Rock Springs uplift (RS)
Gros Ventre Range (GV)	Wind River Range	Sweetwater uplift (SW)
Medicine Bow Mountains	Yellowstone Plateau	

BASINS AND PLAINS

With relics of Pleistocene tundra

Bighorn Basin	Great Divide Basin (GD)	Powder River Basin (PRB)
Carbon Basin (CB)	High Plains (HP)	Shirley Basin (SB)
Casper arch (CA)	Laramie Basin	Washakie plateau (WP)

forming hills and divides as much as 1000 feet above the present basin floors. Because these deposits preceded cutting of the canyons that were subsequently occupied by Bull Lake and Pinedale glaciers, Blackwelder tentatively correlated the Buffalo episode with the Illinoian Stage of North American continental glaciation.



Figure 2. View from Downs Mountian across Grasshopper Glacier to Gannett Glacier (below snow-capped Gannett Peak). Photograph by Finis Mitchell.

Blackwelder named the Bull Lake and Pinedale glaciations from type localities where the prominent end moraines extending several miles beyond the mountain front now enclose Bull Lake on the eastern side of the Wind River Range, and Fremont Lake near the town of Pinedale (**Figures 4** and **5**). Largely from geomorphic evidence he interpreted the moderately subdued Bull Lake moraines, whose outwash plains are on terraces from 80 to 200 feet above present stream levels, as early Wisconsin. He dated the adjacent Pinedale moraines as late Wisconsin because of their prominent knob-and-basin topography (**Figure 6**) and their extension through the Bull Lake end moraines along shallow valleys descending to Pinedale outwash deposits on the flood plains of the adjacent streams.



Figure 3. Dinwoody Canyon and Gannett Peak with glacier and neoglacial moraine (upper center). Photograph by Finis Mitchell.

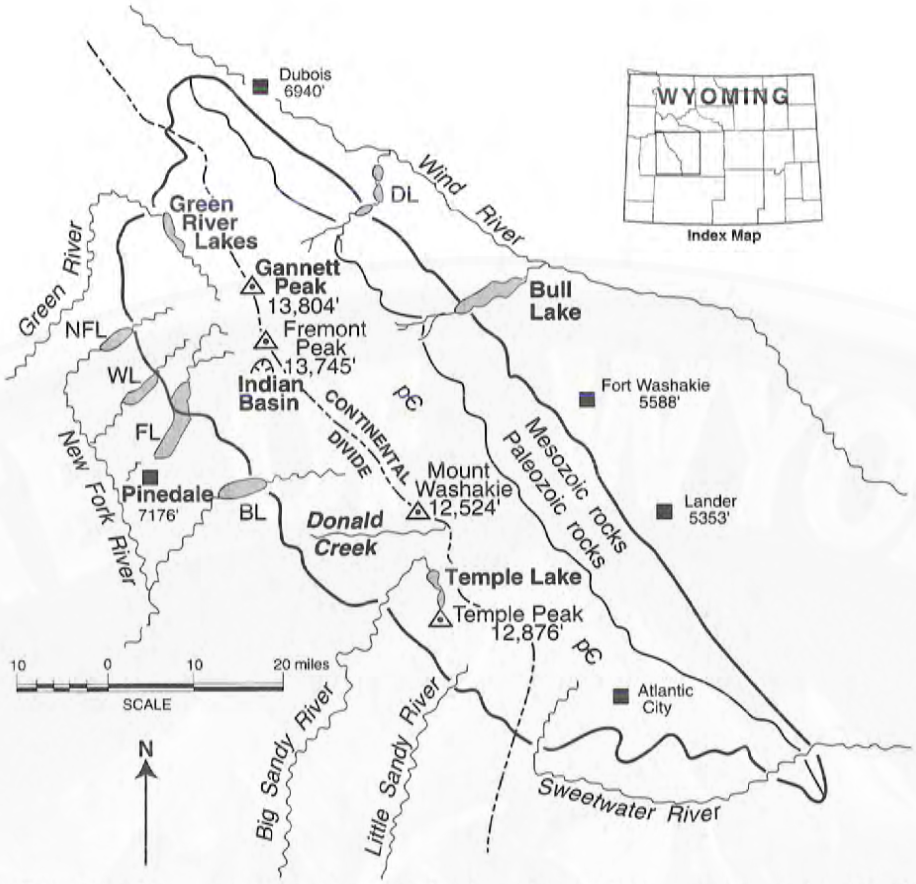


Figure 4. Glacial type localities (boldface), major geologic features, and selected elevations, Wind River Range. Abbreviations for moraine-dammed lakes: BL, Boulder Lake; FL, Fremont Lake; NFL, New Fork Lake; WL, Willow Lake; and DL, Dinwoody Lakes.



Figure 5. Air photograph of Boulder Lake showing Bull Lake (outer) and Pinedale (inner) terminal moraines.



Figure 6. Knob-and-basin topography of Pinedale moraine near Dubois.

Holmes and Moss (1955) were the first to find evidence that Blackwelder's "Buffalo drift" on the southwestern side of the Wind River Range represented at least two glacial episodes separated by an interval of erosion which had produced the deep mountain canyons. Richmond (1962) established the present model for pre-Bull Lake glacial events along the northeastern flank of the Wind River Range where the deposits, which unconformably overlie the Eocene Wind River Formation in the cliffs along Bull Lake, include tills and lake beds containing two soil horizons representing interglacial intervals. Richmond named the pre-canyon deposits Washakie Point (oldest) and Cedar Ridge, and the post-canyon deposits (overlain by Bull Lake moraines) Sacagawea Ridge (**Figure 7**). His attempted correlations with the continental glaciations (Richmond, 1965) have been revised (Richmond, 1986) since the terms Nebraskan and Kansan were abandoned because of the great complexity now recognized in the deposits of the upper Mississippi Valley. The former equivalents in Wyoming are now lumped as pre-Bull Lake (and designated as middle-middle, early-middle, and early Pleistocene deposits).

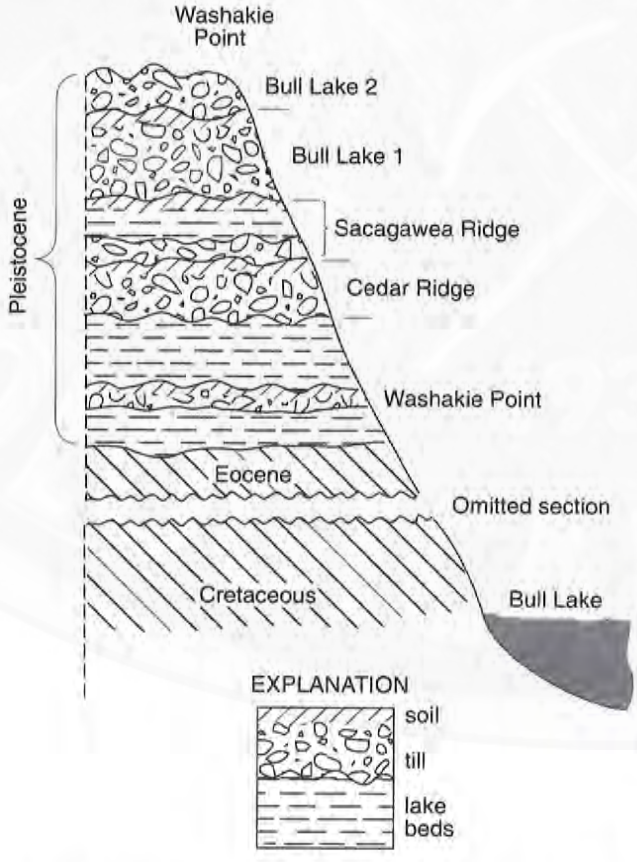


Figure 7. Schematic representation (after Richmond, 1986) of Pleistocene deposits at Washakie Point, along the north side of Bull Lake. The cliff rises 1050 feet above the lake; Pleistocene deposits are 270 feet thick.

The ages in years (absolute ages) for pre-Pinedale glacial deposits are few and are based on long-lived radioactive elements that have large systematic errors when applied to Quaternary events. Richmond and Fullerton (1986) assumed that glaciation began about 1.65 million years before present (y.b.p.), a commonly accepted date for the Pliocene-Pleistocene boundary (derived from a sequence of radiometrically dated deposits in southern Italy). There are no radiometric dates for the Washakie Point tills; however, outwash deposits of Cedar Ridge glaciation have been dated at about 610,000 y.b.p. The age of the Sacagawea Ridge glaciation is known to be greater than 350,000 y.b.p. from dated travertine that overlies Sacagawea outwash in a high terrace along the mountain front a few miles southeast of the town of Dubois (**Figure 4**).

That the Bull Lake glaciation had more than one episode was first suggested by Fryxell (1930) from two sets of moraines and outwash plains in Jackson Hole and later confirmed by Holmes and Moss (1955) at their study site along the southwestern flank of the Wind River Range. Subsequently, Richmond (1962) found three Bull Lake tills, forming some five closely spaced moraine ridges, which he and Murphy (1965 a, b) mapped in great detail at the Bull Lake type locality. A vertical section in the slopes along Bull Lake exposes a paleosol on the underlying Sacagawea Ridge deposits, two disconformities separating the Bull Lake tills, and an upper paleosol beneath overlying Pinedale deposits. There is however no well-developed soil on the surface of Bull Lake moraines exposed to ongoing erosion. The only quantitative date for Bull Lake glaciation in the Wind River Range is a uranium-thorium determination of 160,000 ± 50,000 y.b.p. (Rosholt and others, 1985) from till of the outer Bull Lake moraine near the town of Pinedale. This age fits Pierce and others' (1976) initial confirmation of the Illinoian, rather than Wisconsin, age of Bull Lake glaciation based on potassium-argon dating of associated rhyolite flows, and the thickness of weathering rinds on obsidian pebbles in tills of the West Yellowstone region.

The onset of Pinedale glaciation, estimated at about 35,000 y.b.p. (Richmond, 1985), followed an interval of about 100,000 years during which there is no evidence of glacial ice on the Wyoming basin floors. The Pinedale glacial maximum, marked by the prominent moraines enclosing the major lakes on the basin floors along the Wind River Range, is estimated at about 18,000 y.b.p. During late Pinedale time, the gradual retreat of the ice front was interrupted by minor readvances named from moraines at Green River Lakes, Donald Creek, and Temple Lake near the headwaters of Big Sandy Creek (**Figure 4**). Post-

Pinedale readvances in the high country are represented by early Holocene Indian Basin moraines, late Holocene Audubon moraines (type locality in Colorado), and the Gannett Peak moraines of the 19th Century. Currently some 15 spot samples of the tills and related deposits have been dated with reasonable accuracy, since the Pinedale and Holocene events are within the 50,000-year limit of radiocarbon age determinations.

Permafrost in the basins

During the Pleistocene when glaciers extended beyond the mountain fronts, the unglaciated expanses of Wyoming's basin floors were tundras, underlain by permanently frozen ground marked by ice- and sand-filled wedges forming large polygons of the sort now found in Alaska, northern Canada, and Siberia. The evidence in Wyoming (Mears, 1981, 1987) includes sediment-filled casts of thawed ice wedges ("fossil ice wedges") and relics of the sand-filled types, both forming sizeable polygons (**Figures 8a** and **8b**). They originated from many repeated episodes of thermal contraction, each of which creates directly adjacent narrow vertical cracks 0.2 to 0.3 inch wide (Black, 1969) in the permanently frozen ground.

Sand wedges (**Figure 9a**) form in sparsely vegetated areas where blowing sand directly fills the repeated narrow cracks to create a distinctive vertical fabric. Ice wedges form at better vegetated sites where the sand is stabilized as ice builds up to form a wedge-shaped mass in the constantly recurring cracks. When the climate warms, the ice is replaced by a clastic fill which may contain large cobbles (**Figure 9b**) that fell into the gaps left by melting ice. The host materials cut by wedges include Paleozoic and Mesozoic mudstones, siltstones, shales, and slabby sandstones; Tertiary siltstones and fine-grained pyroclastic rocks; and Pleistocene deposits. Deformation of the host materials, up-bent or down-sagged against the wedge margins, is evident where layering or oriented clasts provide markers but is not discernible in homogeneous hosts. The abundance of wedges that cut bedrock–at 54 of the 85 currently known sites exposed by trenching and bulldozing or natural erosion (Mears, 1987)–reflects the ongoing erosional deepening of the Wyoming basins, which began in late Miocene time (Lillegraven and Ostresh, 1988).

Laboratory study (Nissen and Mears, 1990) shows that the wedge fills are largely poorly sorted silty sands containing abundant quartz and feldspar grains and a few small rock particles. Binocular and scanning electron microscopes show that the larger quartz grains are spherical, like those in present day sand dunes. Unlike the dunes, however, the wedge fills contain many finer particles of conchoidally fractured quartz grains and chips indicating mechanical weathering in a permafrost environment. Although material for radiocarbon dating has not been found in the wedges, they have been dated as Wisconsin at several localities from carbon in directly underlying swamp deposits and from carbon in disconformable overlying deposits which contain archaeological sites.



Figure 8a. Polygons of ice-wedge casts in Pleistocene cobble deposits on Eagle Rocks bench, Laramie Basin.



Figure 8b. Late Pleistocene polygons in Cretaceous shale exposed along U.S. Interstate Highway 80, east of Rawlins.



Figure 9a. Relic sand wedge with the vertical fabric that develops from the successive filling of narrow contraction cracks by sediment during many episodes of intense cold.



Figure 9b. Ice-wedge cast containing a large cobble that fell into the space left when the ice melted.

Pleistocene temperatures

Mean annual air temperatures during the wedge-forming episodes in Wyoming were originally estimated as at least 13°C colder than the present based on the difference between the mean temperature of 8.1°C now recorded at Powell (the warmest site near relic wedges) and the mean temperature 0° to 5°C (Péwé and others, 1969) at the southern limit of active ice wedges near Fairbanks, Alaska. The estimate was based only on air temperatures, which arctic studies report as somewhat colder than those in the ground at ice wedge sites. Lachenbruch (1962) found a 1° to 6°C difference in northern Alaska. Romanovskij (1973), from studies in northern Europe and Siberia, noted that silt cracks at higher temperatures than coarse clastic deposits. Péwé (1983) reported differences in the rapidity, severity, and frequency of temperature changes at active wedge sites in Alaska.

A recent study of ice-wedge casts in coarse clastic hosts on Polecat Bench in the Bighorn Basin, where ground temperatures have been recorded in abandoned oil wells, indicates that the annual average Pleistocene air temperatures may have been at least 17°C colder than those of the present day (Visser and others, 1996). That report includes numerical modeling of various temperatures that would have created contraction cracks extending well down into the permanently frozen ground, beneath the mantle of surficial deposits where summer warming obliterates the upper parts of such cracks. The best of several estimates indicates that the mean annual Pleistocene temperatures were 15.9° to 16.7°C colder than the present day and that their annual range exceeded that of the present day.

Conclusions

During five major Pleistocene episodes when alpine glaciers extended beyond the Wyoming mountain fronts, the snowfall was not much greater than the present but the temperatures were much colder, based on the widespread ice-wedge casts and sand-wedge relics in the high intermontane basin floors and eastern plains. The Wyoming evidence fits some world-wide estimates of late Pleistocene temperature depression. Gates (1976) recorded a -13°C temperature for the Wyoming region on his computer-generated global map simulating the atmospheric circulation 18,000 y.b.p. Washburn (1980) tabulated 30 estimates from worldwide localities for late Pleistocene temperature lowering based on ice-wedge casts. His calculations, adjusted to a common wedge-forming temperature of -5°C, indicate 7° to 20°C lower temperatures, which would include the Wyoming estimate of 17°C lower temperatures.

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The history of industrial mineral development in Wyoming

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Abstract

Industrial minerals have always been produced in Wyoming by its inhabitants. Native Americans quarried stone for various uses including points and tools, they quarried and dug hematite and limonite for pigment (Frey, 1947), and they quarried and shaped steatite into bowls and other implements. They also dug salt from spring deposits, and scraped sodium sulfate from alkali lakes for use in curing hides.

During the western immigration period, there were few permanent inhabitants of what was to become Wyoming. However, those who passed through developed sources of salt, and utilized natural tar deposits for grease.

With the coming of the transcontinental railroad across southern Wyoming in 1867-1869, Wyoming first developed as an agricultural society, with its population spreading north from the railroad. Deposits of clay were found near almost every settlement, and these were used to manufacture brick. Almost every town had a nearby stone quarry, and larger stone quarries were developed near Douglas, Rawlins, and Laramie. Limestone is common in most of Wyoming, and locally was quarried for lime. Local deposits of aggregate were worked, and many towns had a small glass plant, using high-silica sand deposits.

During the period 1890 to 1930, some industrial minerals were produced for regional or national markets. Bentonite was originally developed as a base for cosmetics. A plaster mill, using nearby gypsum deposits, began operating in Laramie. Small scale mines producing vermiculite, mica, graphite, asbestos, and other commodities were developed, usually employing only one or two miners. Production of construction material increased with the increasing population. Phosphate began to be produced in western Wyoming and was calcined and shipped to domestic markets as a soil conditioner. Larger limestone quarries were opened to supply limestone for the refining of sugar from sugar beets. The Jay Em Stone Company was established, producing monuments from a large number of quarries in southeastern Wyoming.

At the end of the 1920s the Great Depression led to closure of most of the small mines, although larger bentonite, limestone, and cement raw material operations continued, with reduced production, until the outbreak of World War II.

World War II saw the increased production of many commodities, especially bentonite. Following the war, trona was accidentally discovered in a core from an oil well, and production began in 1947. Market distances increased, and so did industrial mineral production, through the 50s, 60s, and into the 70s, although the production of phosphate ceased.

In the 1980s, most industrial mineral products in Wyoming saw dramatic declines in their production. Soda ash production from mined trona increased slowly, as international sales grew while domestic consumption fell. Bentonite production fell to 40% of its 1980 maximum, and several plants closed. Production of most materials began to increase again toward the close of the decade, and the 1990s have seen a slow growth in most industrial minerals, particularly bentonite, as new uses for bentonite develop. Recent events include the development of a decorative stone industry, and a dramatic increase in the production of sodium products from mined trona.

Developments before 1867

The area now included in the boundaries of Wyoming before 1867 was principally occupied by Native Americans. It is not known for certain if any Spanish or French explorer or trapper visited Wyoming before 1800, although an authenticated Spanish sword blade was found near Dayton, and de la Vendereye, a French trapper, described what could have been the Bighorn Mountains from a distance (Larson, 1965), although his description could fit the Black Hills of South Dakota equally well. Beginning with the explorations of John Colter in 1807, people of European descent began to explore Wyoming, trapped furs in the area, and moved on, establishing no permanent settlements. Industrial mineral use was limited to immediate needs. During this period, the only industrial mineral products exported from the state were obsidian points from the Yellowstone Plateau (**Figure 1**), some of which have been traced as far east as Ohio.

Native Americans quarried chert, quartzite, and obsidian for points and tools; granite for hammer stones and other tools; steatite (talc) for utensils (bowls and pots) (**Figure 2**); and iron oxides and hydroxides for pigment (**Figure 3**). Salt (NaCl) from natural springs, and sodium sulfate and related materials were used for curing hides (**Figure 1**).

Early trappers used salt and other industrial commodities much as did the Native Americans. The fur trade declined in the 1830s, but the period of western migration began at the end of that decade. Travelers on the Oregon and other trails during the western migration used natural hydrocarbons from tar seeps (**Figure 4**) as axle grease and waterproofing sealants.

Beginning with the establishment of Hartville in 1858(?) as a copper mining camp (see photograph, p. 166), settlements and ranches began to be established

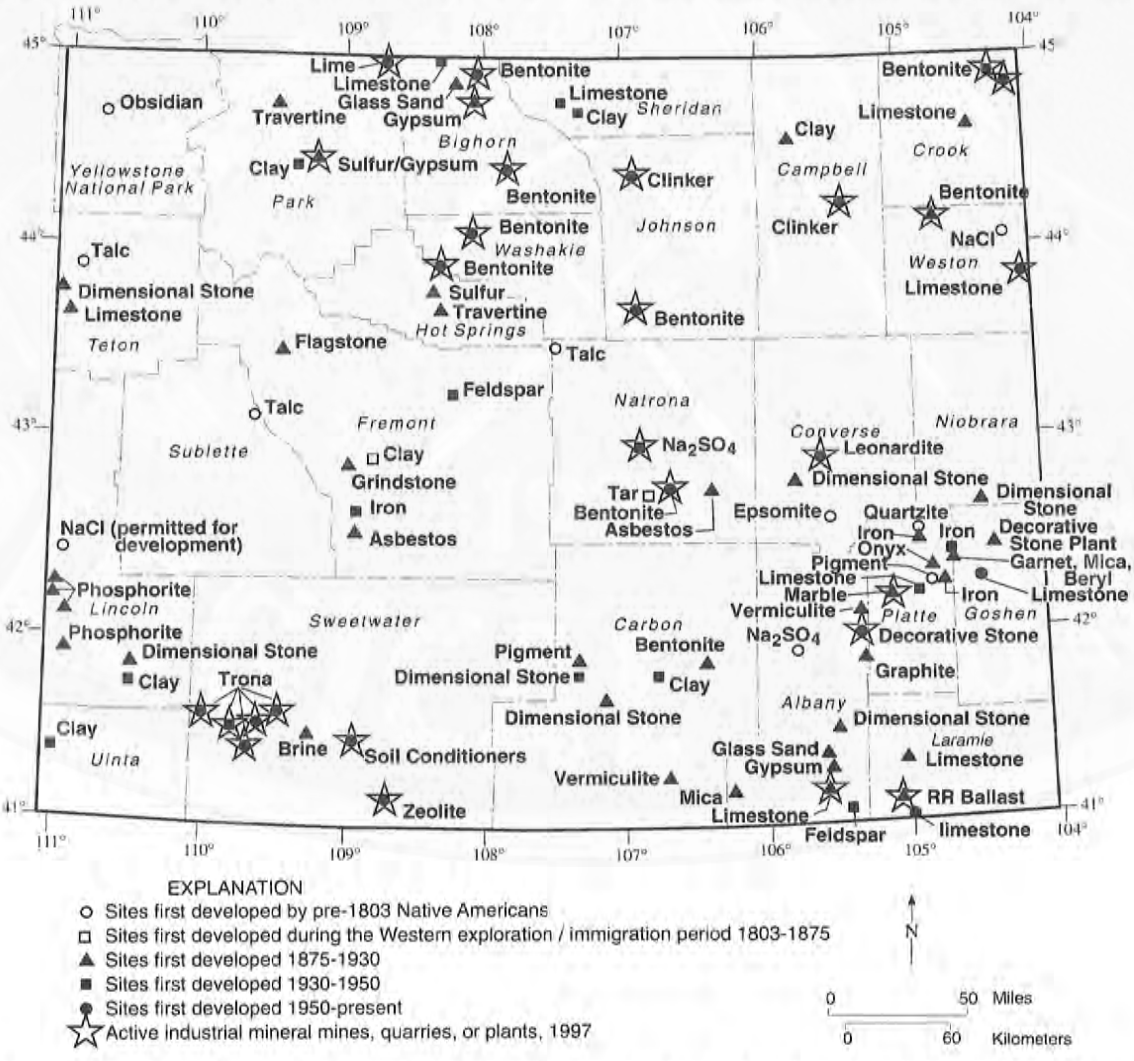


Figure 1. Index map showing historic industrial mineral sources (exclusive of sand and gravel), by date of earliest use. See Harris (1996) for a more detailed map of Wyoming industrial minerals.



Figure 2. Photograph of preColumbian Native American steatite quarry. Backpack in upper right for scale.



Figure 3. Photograph of petroglyph, drawn using native pigments. Field of view is approximately one foot across.

in Wyoming, principally along the valleys of the North Platte and Laramie Rivers in the eastern part of the state. Resourceful settlers used nearby industrial materials for construction. Clay was dug and baked into brick, and limestone was roasted to make lime, which was used in mortar, plaster, whitewash, and numerous other products. Gypsum was used locally in mortar. Local deposits of stone were quarried, shaped, and used in construction, in buildings, sidewalks, rails, and other objects.

Developments from 1867 to 1900

The year 1867 was important in the history of Wyoming because that was the year the transcontinental railroad construction entered the state, linking this remote region to the rest of the country. Not only could products from Wyoming be exported to other locations, but an increasing population linked to agriculture and transportation demanded products that required construction materials and industrial minerals. As an example, many towns established local glass manufacturing for containers, and some localities also saw the production of window-grade glass. Silica-rich sands are found near almost every town, and, if a certain color due to iron impurities could be tolerated, alkali materials (sodium carbonate), feldspar, and other additives found locally could be used in glass manufacture. Likewise, almost every town developed its own clay (for bricks), stone, construction aggregate, and lime (from limestone) source. Larger more specialized industries were established in some towns. Due to the presence of large resources of gypsum, a plaster mill was constructed south of Laramie, and a railroad ballast quarry was established at Granite Cañon, west of Cheyenne.

In addition, a few industrial mineral products were exported from Wyoming. Pigment (goethite) from



Figure 4. Natural tar spring, near the Oregon-California Trail near Casper (photograph by Alan Ver Ploeg, Wyoming State Geological Survey).

Rawlins was exported by the Union Pacific Railroad for use in boxcar and other paints. Building stone from Rawlins (**Figure 5**) was used to construct the Wyoming State Capitol in Cheyenne, the Union Pacific depot in Ogden, Utah, and a few buildings in Denver, Colorado. Bentonite, called wilkinite by the producer, Owyhee Chemical Company of Boise, Idaho, was processed at a small plant in Cheyenne. Owyhee mined its product about two miles west of Medicine Bow, and manufactured a base for cosmetics in the Cheyenne plant. Phosphate from near Sage and Cokeville was mined and distributed to Idaho potato growers. Epsomite-producing plants were established

south of Douglas and north of Rock River (see photograph, p. 144), and several alkali lakes were mined for sodium sulfate along the route of the Union Pacific Railroad (**Figure 1**).

Hardrock industrial mineral deposits were also developed, although hopes for some of these were higher than actual production. These included asbestos deposits at South Pass and especially near Casper, vermiculite and graphite deposits in the Laramie Mountains and Hartville uplift in Platte and Albany Counties, and mica- and feldspar-bearing pegmatites in southeastern and central Wyoming (**Figure 1**). The first shipments of iron ore from Sunrise, near Hartville, to the Colorado Fuel and Iron Company smelter in Pueblo, Colorado, began in 1897 (Osterwald and others, 1966).



Figure 5. "Rawlins Sandstone" quarry four miles southeast of Rawlins is in the Upper Cretaceous Mesaverde Formation.

Developments from 1900 to 1929

This was a period of growth for Wyoming, as well as the rest of the nation. Wyoming did not participate in the boom of the 20s as much as the rest of the nation since, as an agricultural state, industrial growth did not reach everyone (Larson, 1965). Partly because of this, many Wyoming citizens sought out ways to develop nearby mineral deposits. Numerous small operations were developed, and this period of time also saw the larger-scale development of bentonite and phosphate resources. It was during this time that most of Wyoming's industrial mineral deposits were located and described (**Figure 6**), although Wyoming's distance from major population centers precluded their development, due then, as now, to the cost of transportation.



Figure 6. 1920s-era drill rig abandoned sometime in the 1930s in the Laramie Mountains near Marshall, Albany County.

During this period, industrial minerals which were identified as having possible commercial value included the following extensive list (Dietz, 1929): abrasive minerals (garnet, grindstones, pumice and pumicite, and tripoli), asbestos, bentonite, ceramic materials (cement rock [limestone], shale and clay, glass raw materials [silica sand, kyanite, feldspar], and pottery materials [feldspar, refractory clay]), construction materials (sand and gravel, railroad ballast, natural asphalt), fertilizing minerals (phosphate, potash), graphite, gypsum, iron, limestone, mica, mineral pigments, saline and alkaline deposits (sodium chloride, sodium sulphate, sodium carbonate, magnesium sulfate), and sulfur.

Of these, the following were actually being mined, though most were mined in small quantities: asbestos, bentonite, cement rock, clay, dimensional stone, feldspar, garnet, graphite, grindstones, gypsum, iron, limestone, mica, mineral pigments, phosphate, railroad ballast, sand and gravel, silica sand, sodium carbonate, sodium sulfate, and sulfur.

The following were mined in larger quantities and shipped out of state: bentonite; limestone, marl, shale, and gypsum for cement; dimensional stone (monuments and buildings); feldspar, for use in dental appliances and ceramics; iron; limestone, for use in refining beets for sugar; mica; mineral pigments; phosphate; railroad ballast; silica sand and feldspar for glass; sodium sulfate; and sulfur.

Developments from 1929 to 1942

These were the depression years, and Wyoming's industrial mineral production declined along with that of the rest of the nation. Significantly, many small mines closed. Most of the smaller mines and prospects

throughout Wyoming were abandoned at this time (**Figure 6**). The only industrial mineral commodities that survived the depression were the bentonite, cement, sugar rock, and phosphate industries although construction materials (brick clay, construction aggregate, railroad ballast, and stone) continued to be produced as well.

Developments from 1942 to 1948

World War II brought about a revival of large-scale industrial mineral production, along with the testing of new sources for industrial minerals. A major inventory of minerals in the state was financed by the Federal government, the Bureau of Mines, other Federal agencies, and the Wyoming State Geological Survey. However, it was the accidental discovery of trona (that resulted from the prewar increase in oil exploration) that was to prove the most significant discovery of an industrial mineral in Wyoming to date.

Although sodium carbonate brines had been identified in wells at Green River since 1896, it wasn't until 1938 when an oil exploration well was drilled in sec. 2, T18N, R110W, that extensive beds of sodium carbonate-bicarbonate, known as trona, were discovered in the subsurface (Greene, 1941). Before the United States entered World War II, four holes were drilled in the area, and an area of trona resources was identified. Exploration and development ceased during the war, but in 1947, Westvaco Chlorine Products Company (now FMC) sunk a shaft to the trona deposit, and in 1948 began underground mining and calcining trona, producing soda ash (Thomas, 1966).

Bentonite and phosphate production reached record levels, as did the major sources of mineral wealth in Wyoming: oil, coal, and iron ore. Gypsum deposits were developed near Cody, and raw gypsum was calcined and shipped out of state.

In an effort to locate a domestic source of aluminum, a research facility at Laramie was established by the U.S. Bureau of Mines and the War Materials Board to study ways to process the anorthosite deposit in the Laramie Mountains for a source of aluminum. Other uses for the anorthosite were also considered, including refractory materials and nonconducting materials.

Developments from 1948 to 1980

These were the boom years for Wyoming industrial mineral production, although production did not increase steadily every year (**Figures 7**, **8**, and **9**). The development of trona (**Figure 7**), bentonite (**Figure 8**), and gypsum (**Figure 9**) led to the growth of major industries in the state. This time period also saw the intermittent mining of clay, feldspar, decorative and dimensional stone (primarily at the University of Wyoming quarry in the 1960s, **Figure 10**), and sodium sulfate. The production of limestone for sugar rock declined, and the mining of phosphate in Lincoln County ceased in 1977 (**Figure 11**).

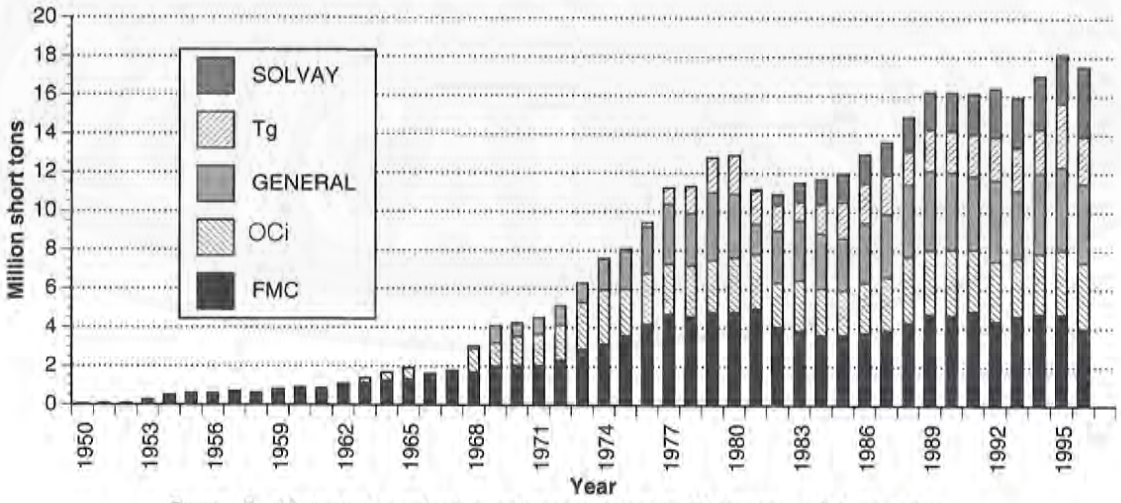


Figure 7. Historical production of mined trona in Wyoming, 1950 to 1996.

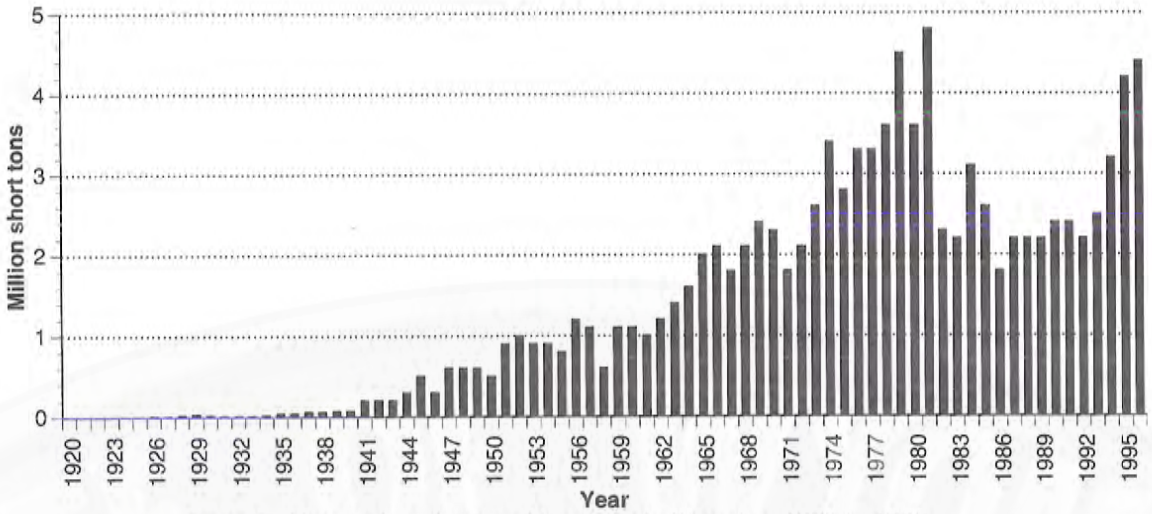


Figure 8. Historical production of bentonite in Wyoming, 1920 to 1996.

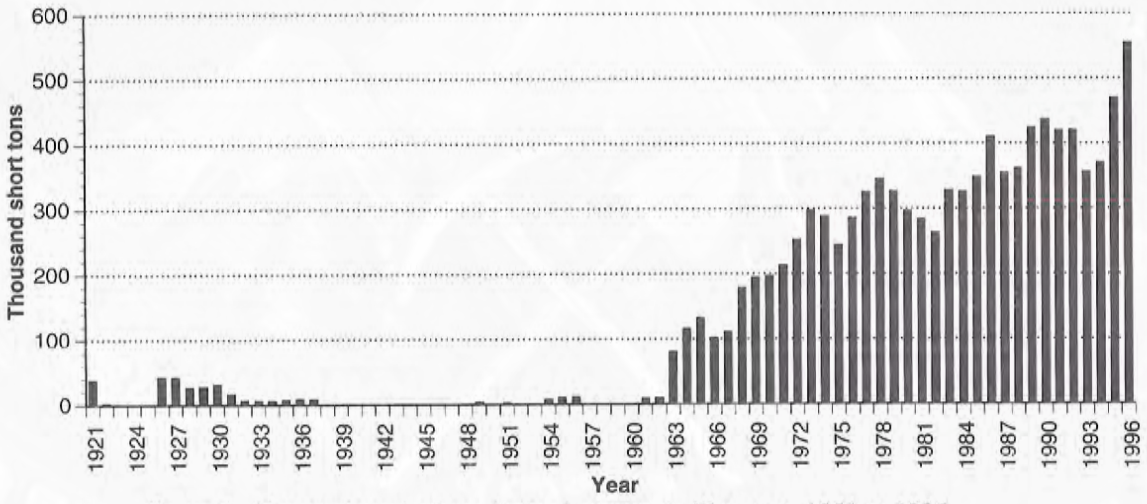


Figure 9. Historical production of mined gypsum in Wyoming, 1921 to 1996.



Figure 10. University of Wyoming dimensional stone quarry north of Laramie. Photograph taken in 1984 when the gantry (to left) was in place. The gantry has since been removed.

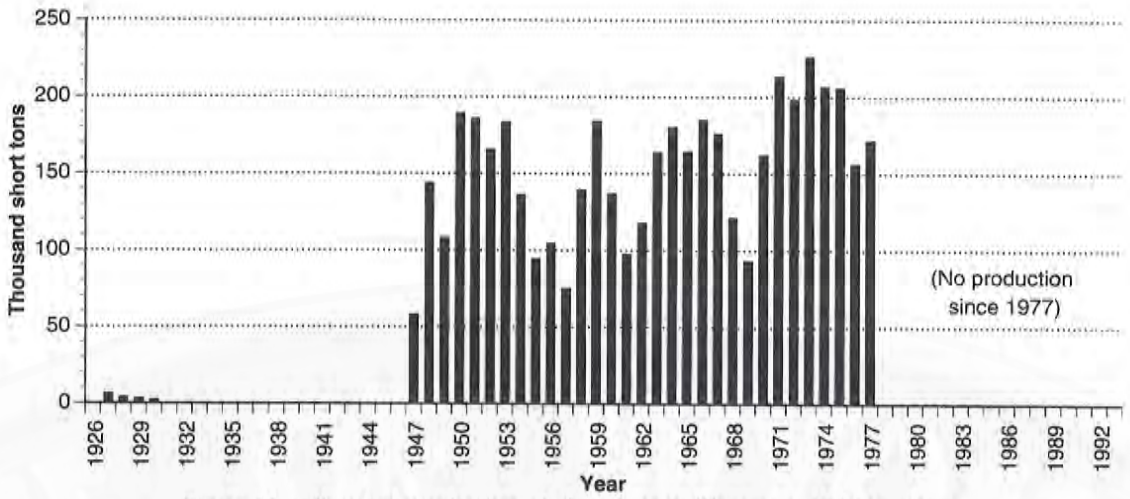


Figure 11. Historical production of phosphate in Wyoming, 1926 to 1977.

Developments from 1980 to the present

The production of the major industrial mineral products of Wyoming peaked in the years 1979 to 1982 (**Figures 7**, **8**, and 9). The years following were a period of readjustment, with renewed growth from 1991 to 1993. Trona production declined mainly because of the decline in the container glass demand as plastic bottles replaced glass, but production increased later due to increased exports to the Far East, North and South America, and, to a lesser extent, to Europe. Bentonite production fell as domestic oil exploration entered a depression, but began to increase as other uses, such as environmental containment barriers, taconite pelletizing, and, most notably, kitty litter, were marketed. Gypsum production declined with the construction slump of the 80s, but increased again in the 90s. In 1996, the two wallboard plants in Wyoming were producing at capacity.

A decorative stone (black granite) quarry opened in 1993 (**Figure 12**) and has steadily increased its production since then. A few small companies in Wyoming produce decorative stone, fieldstone, sandstone, marble, and moss rock, which is sold to national markets. The production of white marble aggregate and fines at Georgia Marble's Wheatland operation has increased.

Meridian Aggregates operates an aggregate quarry west of Cheyenne (see photograph, p. 106) that formerly supplied only railroad ballast. Recently, Meridian has sold a quartzofeldspathic gneiss aggregate from this quarry for other commercial uses. Meridian also supplied 1 million short tons of aggregate that was used in construction of the new Denver International Airport. That amount was 10% of the total amount of aggregate used in the airport's construction. Meridian's product is also one of the few materials in existing production that can meet the requirements for superpave (Bill Langer, U.S. Geological Survey, personal communication, 1996).

Other construction aggregate producers supply material to construction projects statewide, and Guernsey Stone, at Guernsey, Wyoming, supplies aggregate to projects throughout western Nebraska, an area devoid of good quality crushed stone. Mountain Cement added a second kiln in 1995-1996, and is increasing production of cement. High-grade feldspar is mined in the summer months on Casper Mountain by Pacer Corporation, shipped to Custer, South Dakota, and used to make high-purity ceramic glazes.

Future industrial mineral possibilities in Wyoming

Wyoming's production of industrial minerals is, as it always has been, transportation dependent. There is not much of a market within the least populated state in the country, and regional markets (Denver, Colorado; Salt Lake City and the Wasatch front area of Utah; the eastern Snake River Plain of Idaho; Billings, Montana; Rapid City and the Black Hills of South Dakota; and western Nebraska) are also small. Wyoming's principal industrial mineral production is of specialized materials found only in Wyoming (trona, high quality feldspar, specialty decorative stone, and bentonite), or found in Wyoming closest to major markets (gypsum/wallboard, cement, limestone). The production of these products is expected to continue at a

rate determined principally by the world economy and import restrictions. The small local demand for materials like sand and gravel will support a small local industry, and growth in the Denver-Colorado Front Range urban corridor will require increased use of Wyoming construction materials such as granite aggregate and limestone.

Minerals found in Wyoming which hold the possibility of becoming specialty sources include zeolites (see photograph, p. 268) and diatomite. Innovative production methods may favor glass production in Wyoming, and, since all of the raw materials necessary for the manufacture of glass (silica sand [**Figure 13**], soda ash, and feldspar) are found in the state, glass may be produced in the future. A growing decorative stone industry may be the result of changes in the world economy, as long as Wyoming's economic advantages continue.

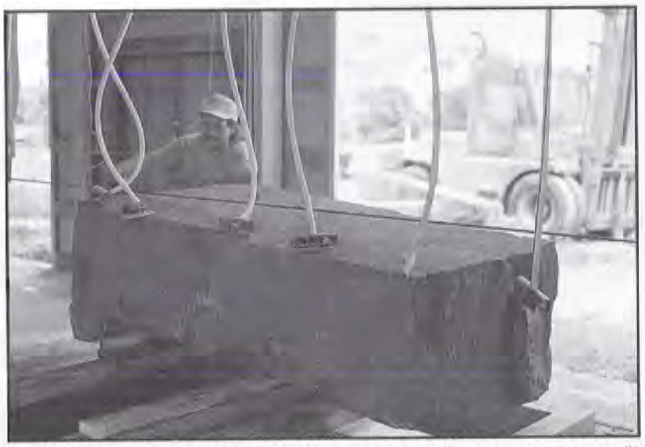


Figure 12. Sunrise Stone's black granite ("Wyoming Raven") plant, Platte County, Wyoming.



Figure 13. Silica sand deposit 30 miles north of Laramie.

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Sodium carbonate resources of the Green River Formation

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Abstract

Marly lacustrine rocks of the Eocene Green River Formation, which occupy parts of several basins in Wyoming, Colorado, and Utah, contain world-class resources of oil shale and sodium carbonate minerals. The formation contains the world's largest known resource of natural sodium carbonate as bedded trona ($Na_2CO_3 \cdot NaHCO_3 \cdot 2H_20$) in southwestern Wyoming as well as the second largest known resource of sodium carbonate as bedded and nodular nahcolite ($NaHCO_3$) in northwestern Colorado. Five companies currently mine three beds of trona in Wyoming and one company mines a bed of nahcolite in Colorado.

Other sodium carbonate minerals in the Green River Formation that may have future economic potential include dawsonite [$NaAl(OH)_2CO_3$], eitelite ($Na_2CO_3 \cdot MgCO_3$), and shortite ($Na_2CO_3 \cdot 2CaCO_3$). Several sodium carbonate brines in Utah and Wyoming may have economic value for soda ash and for water-soluble organic acids.

Introduction

The lacustrine Green River Formation of Eocene age, which occupies several sedimentary basins in Wyoming, Colorado, and Utah (**Figure 1**), has long been known for its oil shale resources but less so for its enormous deposits of sodium carbonate minerals. Large deposits of sodium carbonate salts occur in the three-state area. These deposits include relatively pure beds of trona, nahcolite, and eitelite; nonbedded aggregates, nodules, and crystals of nahcolite; microscopic crystals of dawsonite in oil shale; veinlets and crystals of shortite; and brines of which some are highly enriched in organic acids. Five sodium carbonate occurrences including two in Utah, one in Colorado, and two in Wyoming, are being mined or have potential economic value.

The occurrences include deposits of bedded trona in the Green River Basin, southwestern Wyoming; deposits of nahcolite and dawsonite commingled with high-grade oil shale in the Piceance Creek Basin, northwestern Colorado; and bedded sodium carbonate salts near the town of Duchesne in the Uinta Basin, northeastern Utah. The formation also contains two sodium carbonate brines that may have potential economic value. One of these brines occurs in a confined aquifer, the Birds-nest zone, in the eastern Uinta Basin about 30 miles south and east of Vernal, Utah. The other sodium carbonate brine–noteworthy for its usually high content of soluble organic matter–underlies several townships at shallow depths near the communities of Farson and Eden in the northeastern part of the Green River Basin, Wyoming (**Figure 1**).

In this paper, the geologic setting of the Green River Formation is presented and the economic geology of each of the five occurrences of sodium carbonate and their potential for commercial development is described.

Geologic setting

Lacustrine sediments of the Eocene Green River Formation were deposited in two large lakes that occupied several lacustrine sedimentary-structural basins in Colorado, Wyoming, and Utah. These basins include the Green River, Washakie, and Sand Wash Basins in southwestern Wyoming and northwestern Colorado, the Piceance Creek Basin in northwestern Colorado, and the Uinta Basin in northeastern Utah (**Figure 1**).

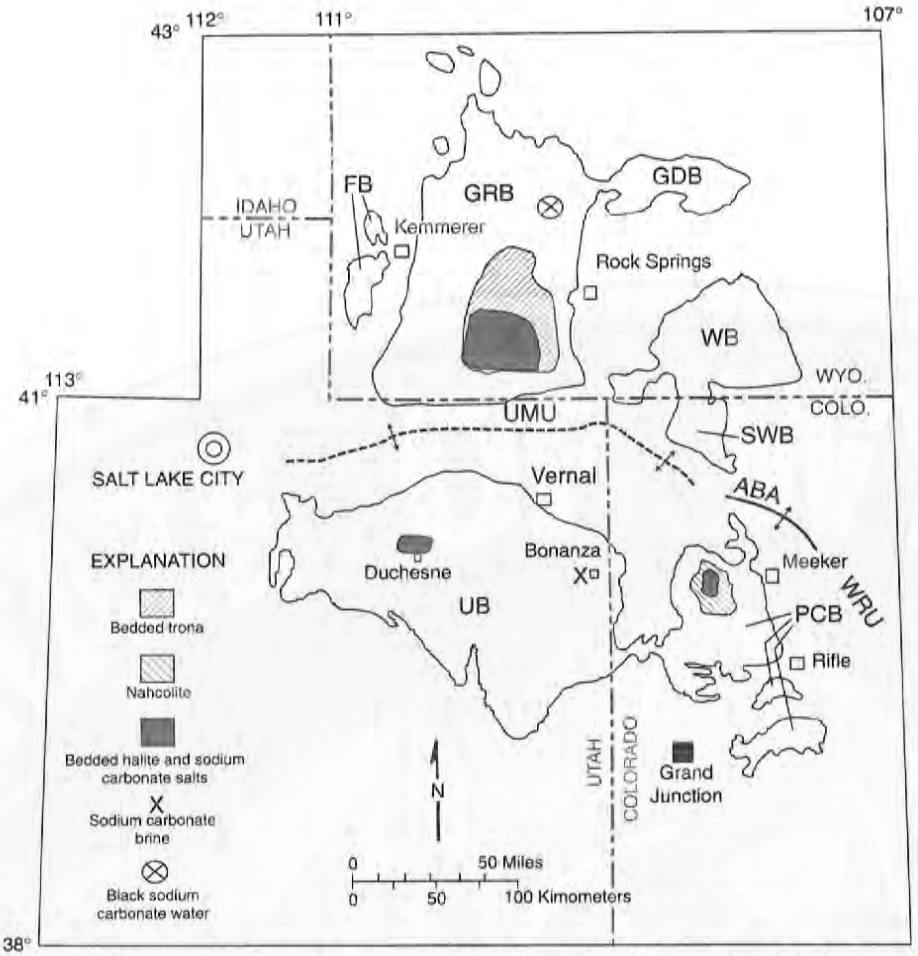


Figure 1. Map showing the areal distribution of the Green River Formation and the locations of five sodium carbonate deposits in Utah, Colorado, and Wyoming. Abbreviations are: ABA, Axial Basin anticline; FB, Fossil Basin; GRB, Green River Basin; PCB, Piceance Creek Basin; SWB, Sand Wash Basin; UB, Uinta Basin; UMU, Uinta Mountain uplift; WB, Washakie Basin.

Lake Gosiute is the northern Eocene lake that occupied the Green River, Washakie, and Sand Wash Basins. The southern lake, Lake Uinta, occupied the Uinta and Piceance Creek Basins in Utah and Colorado. The Uinta Mountain uplift, an east-west positive structural element, separated the two lakes, although they may have been connected during part of their existence on the east side of the uplift. At its maximum extent, Lake Uinta covered about 22,000 square miles (57,000 km²), and Lake Gosiute, about 15,000 square miles (38,850 km²). Together, the two Eocene lakes were somewhat larger than present-day Lake Erie and Lake Michigan combined. The drainage basins of the two lakes were probably double or more in size than the areas of the lakes at their maximum extent (**Figure 2**).

With the exception of tuffaceous sediments, the drainage basins of the Eocene lakes were the source of most of the Green River sediments. Volcanic tuffs were deposited directly into the lake and tuffaceous sediments also were washed into the lake from the surrounding drainage basin. Tuffaceous debris amounts to roughly 5% of the total sediments in the formation.

The glassy tuffaceous sediments readily decomposed in the alkaline lake waters and no doubt supplied some of the dissolved sodium and other cations that comprise some of the authigenic carbonate minerals found in the formation.

The Green River Formation consists of a variety of lacustrine sedimentary rocks including sandstone, claystone, shale, dolomitic marlstone, oil shale, and numerous, generally thin, beds of volcanic tuff that were deposited in the two lakes. During much of Green River time, the lakes were closed with no outlets. During arid periods, the lakes became sufficiently saline to precipitate large quantities of sodium salts, either directly from the lake waters or interstitially from briny waters within the sediments. The most abundant of these salts include trona, nahcolite, shortite, dawsonite, halite, and less commonly, eitelite, and wegscheiderite.

Typically, the rocks grade from fine-grained sandstone, mudstone, claystone, and shale in shoreward areas of the lakes into carbonate-bearing rocks enriched in organic matter toward basin center. The lateral transition from siliciclastic to carbonate-enriched rocks is also reflected in vertical sequences of the Green River Formation in the Green River, Piceance Creek, and Uinta Basins. This indicates increasing salinity of the alkaline lake waters as divalent cations of calcium and magnesium were precipitated as calcite, mg-calcite, and dolomite, while dissolved sodium, bicarbonate, and chloride continued to accumulate in waters in the deeper parts of the basins. Sulfate ion was essentially completely reduced by bacterial redox processes whereby bicarbonate was formed in the ratio of two moles of bicarbonate for each mole of sulfate that was reduced. Hydrolysis of detrital silicate minerals and glassy volcanic ash contributed addi-

tional dissolved bicarbonate, sodium, and other ions to the lake waters.

During their existence, the Green River lakes fluctuated widely in size in response to wet and dry climatic cycles. During three major dry periods, the lakes contracted sufficiently to precipitate sodium salts in each basin of the three-state area. The first and most important dry cycle was during Wilkins Peak and Parachute Creek time when the vast deposits of trona and nahcolite were deposited in the Green River and Piceance Creek Basins, respectively. During two later dry periods, nahcolite in the Birds-nest zone was deposited in nodules and possibly some thin beds in the eastern Uinta Basin, Utah. Finally, during a late dry period, bedded sodium salts were deposited in the Uinta Basin near Duchesne, Utah.

After deposition and lithification of the Green River sediments and their associated sodium carbonate deposits, ground waters dissolved some of these sodium minerals as evidenced by the presence of solution breccias and solution cavities in drill cores recovered from exploratory wells drilled into the formation. In at least two areas, the Green River Formation contains sodium carbonate brines of unknown, but possibly exploitable, size that were probably formed by dissolution of sodium minerals by ground waters.

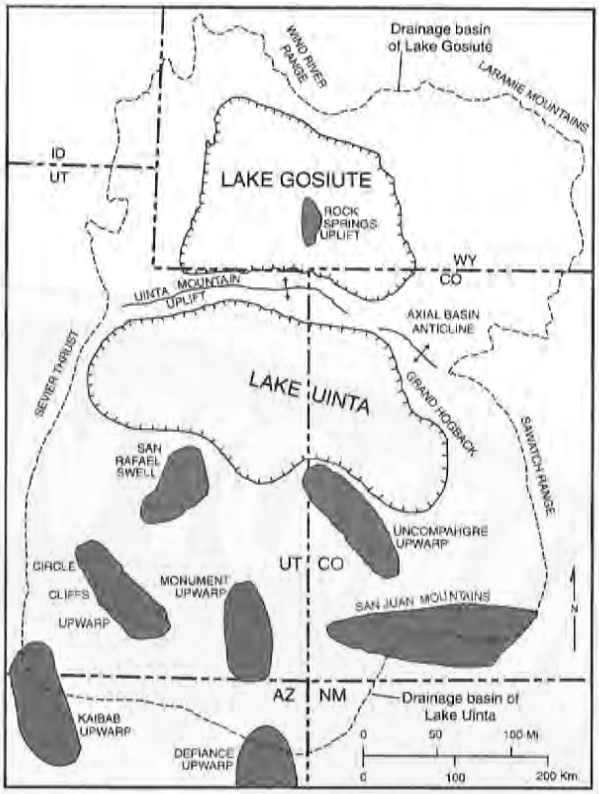


Figure 2. Paleogeographic map of the Eocene Green River lakes at their maximum extents and their approximate drainage basins.

Duchesne sodium carbonate deposit

The occurrence of abundant crystal molds of a then unknown salt in exposures of marly sedimentary rocks of the Green River Formation in the eastern Uinta Basin and in the Piceance Creek Basin, lead W.H. Bradley (1931, plate 3) to define several "saline phases" in the formation. One of these "phases" included part of a sequence of marlstone and shale in the upper part of the formation in Indian Canyon, near Duchesne, Utah (**Figure 1**), which he designated the "barren and saline facies." Identification of one of the salts in the upper part of the Green River Formation in the Uinta Basin was first made by Erickson (1952) who reported the occurrence of shortite in drill core from the Sun Oil Company core hole located in sec. 22, T9S, R20E, Salt Lake Meridian, Uintah County, Utah, between the depths of 1975 and 2056 feet. Subsequently, Milton and others (1954) and Milton (1957) reported the occurrence of shortite and other authigenic minerals from bore holes in the Uinta Basin, including the new mineral, eitelite ($Na_2CO_3 \cdot MgCO_3$). J.J. Fahey (Dane, 1955) identified shortite and nahcolite, possibly bedded, in cuttings from the Joseph Smith oil and gas test well 1 in sec. 16, T3S, R5W, a few miles north of Duchesne (Well No. 2, **Figures 3** and **4**).

In 1970, Mapco Oil Company drilled the No. 1 Shrine Hospital well in the NE SE, sec. 10, T3S, R5W, about 4 miles north of Duchesne (Well No. 3, **Figures 3** and **4**). Two sequences were cored between the depths of 4140 to 4147 and 4165 to 4252 feet (1262 to 1264 and 1269 to 1296 m). The lower core revealed three beds 2, 6, and 12 feet (0.6, 1.8, and 3.7 m) thick of halite (60 to 80%) mixed with varied amounts of wegscheiderite (20 to 40%) and minor amounts of marlstone. The lower core also revealed an additional six thin beds of mixed nahcolite, trona, wegscheiderite, and eitelite ranging from 0.3 to 1.5 feet (0.1 to 0.5 m) in thickness (Dyni and others, 1985).

In 1973, the Duchesne deposit was further tested by Marine Minerals core hole 2 drilled in the SE NW sec. 34, T3S, R5W, about one mile west of Duchesne (**Figures 4** and **5**). The bore hole was continuously cored from 50 to 2857 feet (15 to 871 m) below the surface. The top of a sequence of eitelite-bearing marly rocks, 20.3 feet (6.2 m) thick, was reached in this well at a depth of 2742.7 feet (836 m) (**Figure 5**). A 4.8-foot (1.5-m)-thick bed containing about 75% eitelite was found at a depth of 2757.5 feet (840 m).

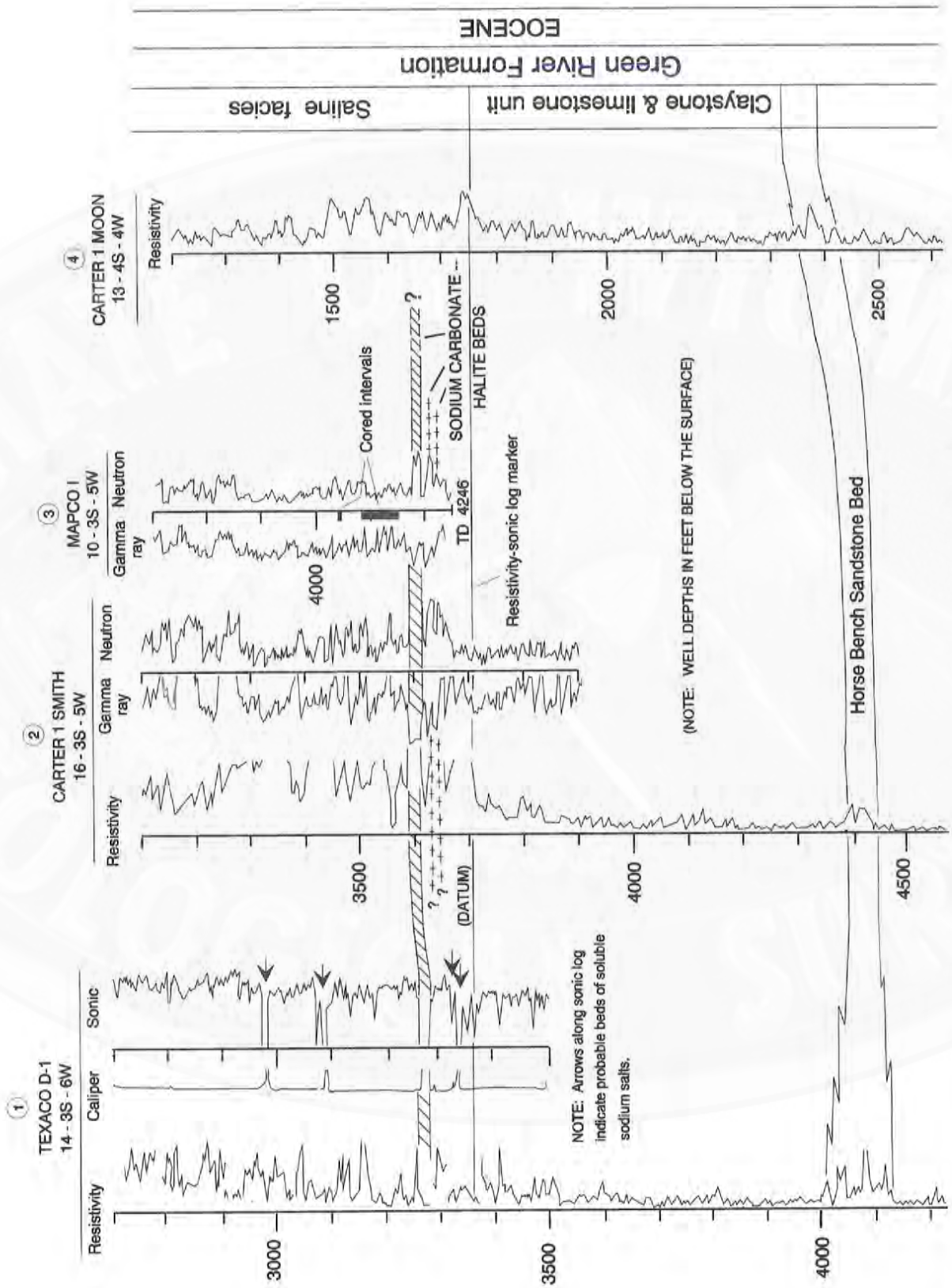


Figure 3. Stratigraphic cross section of the saline facies and underlying rocks in the upper part of the Green River Formation near Duchesne, Utah. Circled numbers above each drill hole are keyed to the line of stratigraphic section shown in **Figure 4**. *Arrows point to possible evaporite beds.*

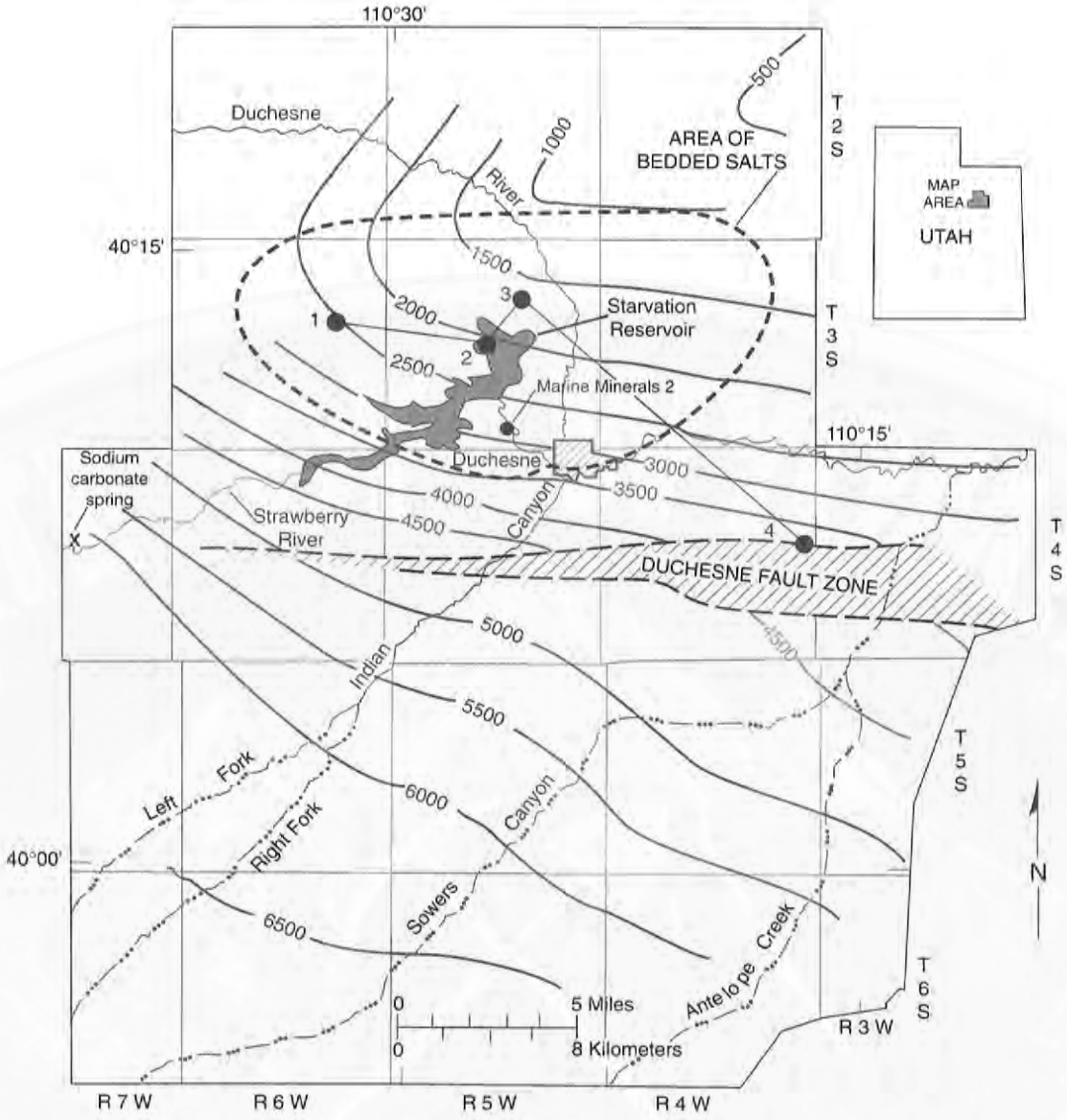


Figure 4. Map showing approximate limits of bedded sodium salts in the upper saline facies (heavy dashed line) near Duchesne, Utah, and line of stratigraphic section shown on ***Figure 3****. Structure contours, in feet, drawn on a resistivity-sonic log datum near the base of the saline facies (see* ***Figure 3****).*

The net thickness of eitelite in the 20.3-foot-thick sequence of eitelitic rocks is about 10.7 feet (3.3 m). The 4.5-foot (1.4-m)-thick tuff in core hole 2 (**Figure 5**) is the same tuff (6 feet thick) identified as unit 11 in the Mapco No. 1 Shrine Hospital core hole (Dyni and others, 1985, figure 4). Shortitic marlstone and other thin beds of mixed sodium carbonate salts and halite were also noted in core hole 2. The core from this well is stored at the U.S. Geological Survey's Core Research Center, Lakewood, Colorado.

The bedded salts at Duchesne occur in the lower part of the saline facies of the Green River Formation as defined by Bradley (1931, plate 9). In Indian Canyon south of Duchesne, the saline facies is about 1100 feet (335 m) thick and its base is marked by a tuff 3 feet (0.9 m) thick (identified as a coarse-grained tuffaceous sandstone by Bradley, 1931, plate 9, base of column D). This tuff may the same as the thick tuff bed noted in Mapco Shrine Hospital 1 well and in Marine Minerals core hole 2 (**Figure 5**).

The saline facies in Indian Canyon consists of fine-grained marly rocks including cherty marlstone, claystone, "mealstone," dolomite, limestone, and minor amounts of oil shale, sandstone, siltstone, and tuff. Mealstone is a name given to a marly rock containing abundant euhedral calcite crystals pseudmorphic after shortite. Water-soluble saline minerals appear to be especially abundant in the lower two-thirds of the saline facies. Although bedded salts were encountered in wells at Duchesne, no solution breccias marking the

former presence of such beds were found on the outcrop in Indian Canyon (Dyni and others, 1985). Bryant and others (1989, plate 1) have mapped the saline facies as far west as Ts8-9S, R6E, along the Wasatch-Uinta County line about 45 miles west of Duchesne in about the same area where Dyni and others (1985, p. 53-54) noted crystal cavities in the unit.

Underlying the saline facies is a unit, about 460 to 700 feet (140 to 213 m) thick, of predominantly claystone, subordinate limestone, and numerous thin tuffs, which is devoid of salts. The claystone and limestone unit is underlain by the Horse Bench Sandstone Bed (**Figure 3**).

The bedded salts at Duchesne can be traced on geophysical logs into other wells drilled in the area. Several beds of mixed halite and sodium carbonate salts are shown on the cross section in **Figure 3**. The most prominent of these beds, unit 21 of Dyni and others (1985, figure 4), ranges from 2 to 19 feet (0.6 to 5.8 m) in thickness; it may not be present in the Mapco core hole 2. Other probable beds of sodium salts are also indicated on geophysical logs in **Figure 3**. Bedded salts underlie an area of about 86 square miles (223 km²) just north of Duchesne (Dyni and others, 1985, figure 1), but the area could be larger (**Figure 4**).

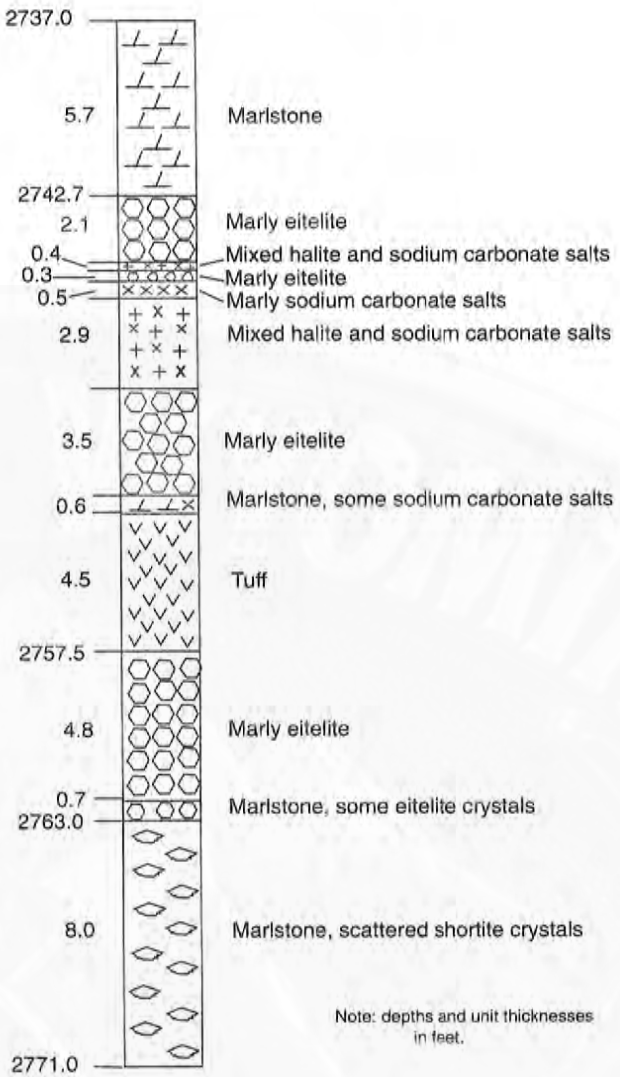


Figure 5. Generalized lithologic log of the bedded sodium carbonate salts and halite in the Marine Minerals core hole 2 in the SE NW sec. 34, T3S, R5W, Duchesne County, Utah (unpublished data from Irvin Nielsen).

Bird's-nest brine field

During a brief, but strongly regressive phase of Lake Uinta during late Green River time, the Horse Bench Sandstone Bed was deposited over a wide area in the south-central part of Lake Uinta. The bed grades toward basin center to the north into open-lacustrine marly kerogenous rocks. During this regressive phase, the remaining lake waters in the deeper part of the Uinta Basin became concentrated in salts, chiefly sodium bicarbonate with probable lesser amounts of sodium chloride. During early diagenesis, when the interstitial salty waters in the marly sediments were being expelled, nahcolite was precipitated within the marly sediments as scattered crystals, nodules, and crystalline aggregates. A few thin beds of nahcolite may have also been deposited. On the outcrop, the nahcolite-bearing sequence was informally named the "Bird's-nest zone," in reference to the numerous cavities left by dissolution of the nahcolite nodules by surface waters which resemble swallow nests (Cashion, 1967, p. 17).

Percolating ground waters have dissolved the nahcolite forming an aquifer in the upper part of the Green River Formation in an area of about 250 square miles (647 km²) in parts of 14 townships (Ts8-11S, Rs22-25W, Uintah County (**Figure 6**), about 30 miles south and east of Vernal, Utah (Lindskov and Kimball, 1984, figures 13 and 20). The full extent of the aquifer has not yet been determined. In the southern part of the area in T10S, R24E, analyses from wells drilled on Federal lease tracts Ua and Ub indicate that the aquifer is slightly to moderately saline, ranging from 1000 to 4800 milligrams/liter of total dissolved solids (**Figure 6**). However, much higher salinities were encountered in the aquifer in two core holes drilled by the U.S. Geological Survey some 6 to 10 miles (10 to 16 km) to the northwest in T9S, Rs22 and 23E (**Figure 6**). Chemical analyses of the brine in the aquifer in the two wells listed in **Table 1** are from Scott and Pantea (1982a, 1982b).

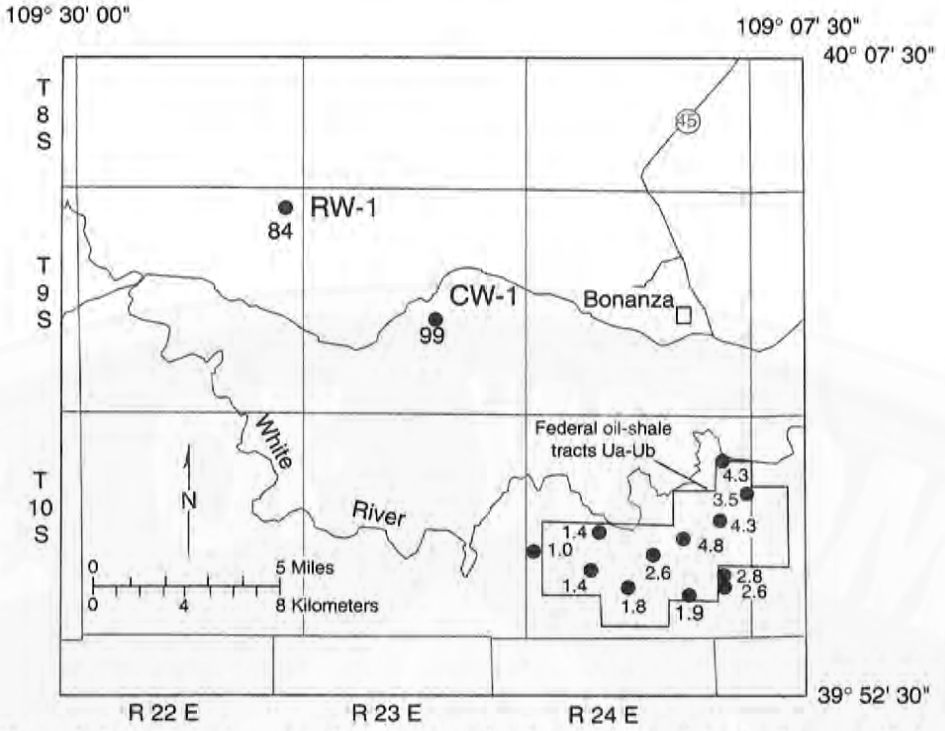


Figure 6. Locations of some wells drilled into the Bird's-nest aquifer near Bonanza, Utah. Numbers indicate total dissolved solids in grams/liter (from Lindskov and Kimball, 1984; Scott and Pantea, 1982a, 1982b).

Table 1. Chemical analyses of the Bird's-nest aquifer in U.S. Geological Survey core holes Coyote Wash 1 and Red Wash 1. See Scott and Pantea (1982a, 1982b) for additional analytical details. Thicknesses and depths in feet (meters); yields in gallons/minute (liters/minute).

	Coyote Wash 1	**Red Wash 1**
Location	SE NE Sec. 22, T9S, R23E	SE NW Sec. 1, T9S, R22E
Thickness of aquifer	65 (19.8)	80 (24.4)
Depth to top of aquifer	1739 (530)	2021 (616)
Sample depth	1746 (532)	2021 (616)
Rock matrix	Oil shale	Oil shale
Yield	700 (2650)	700-900 (2650-3400)
Ion[1]	**Chemical composition (milligrams/liter)**	
Cl^-	3100	650
F^-	26	450
SO_4^{--}	7400	3500
$(HCO_3^-/CO_3^{--})^2$	(13,130)	(13,890)
B	180	240
Na^+	34,000	41,000
K^+	39	60
Ca^{++}	3	<1
Mg^{++}	1	0.2
SiO_2	<2	<2
Organic carbon, dissolved	na	60
pH	9.5	9.5
Alkalinity (as $CaCO_3$)	79,300	105,000
(Total dissolved solids)[2]	83,900	98,600

[1]Other ions found in concentrations of less than 1 mg/l: As, Ba, Cr, Fe, Li, Mn, Hg, and Zn.
[2] The sum of CO_3+HCO_3 and total dissolved solids, which were not determined in the original analyses, were estimated by Scott R. Charlton, University of Colorado, using the U.S. Geological Survey WATC computer program.

Further exploration of the Bird's-nest brine field in the vicinity of the Coyote Wash and Red Wash core holes in T9S, Rs22 and 23E and surrounding area (**Figure 6**) may reveal exploitable reserves of sodium carbonate brine. Dissolved components in amounts that might be troublesome in developing this resource include chloride, sulfate, and fluoride.

Colorado nahcolite deposits

A large resource of nahcolite is commingled with oil shale in the Parachute Creek Member of the Green River Formation in the northern part of the Piceance Creek Basin in northwestern Colorado. The geology of the deposit is well documented by numerous studies, although details of the stratigraphic distribution and grade of nahcolite in selected zones and beds need to be elucidated.

In the Piceance Creek Basin, the Green River Formation is divided, in ascending order, into the Anvil Points, Garden Gulch, and Parachute Creek Members. The Green River is overlain by sandstone and siltstone of the Uinta Formation (**Figure 7**). Most of the significant deposits of oil shale and all of the sodium carbonate resources are found in the Parachute Creek Member.

In the deeper northern part of the Piceance Creek Basin, the Parachute Creek Member contains a saline facies which includes beds of nahcolite and halite, scattered crystals, nodules, and coarse-crystalline aggregates of nahcolite commingled with high-grade oil shale. Also included in the saline facies is dawsonite which occurs as microscopic crystals that comprise part of the microcrystalline matrix of the oil shale.

The top of the nahcolite-bearing oil shale is a dissolution surface above which ground waters have dissolved nahcolite and halite forming a "leached zone" of vuggy brecciated oil shale several hundred feet thick. The leached zone extends upward into the lower part of the Mahogany oil-shale zone, a unit of high-grade oil shale traceable throughout the basin. The top of the leached zone is defined by the stratigraphically highest occurrence of solution breccia that marks the former presence of a bed of nahcolite or mixed halite and nahcolite (Dyni,1981, plates 1 and 2). This top can be considered the top of the original bedded saline facies in the Parachute Creek Member, although crystal cavities and vugs are found scattered through the upper part of the Parachute Creek Member above the Mahogany oil-shale zone (**Figure 7**).

The maximum thickness of nahcolite-bearing oil shale is reached in Ts1N-2S, and Rs97-98W (**Figure 8**). Here, the dissolution surface lies at depths of about 1400 to 1900 feet (427 to 579 m) below ground level (**Figure 7**). Below the dissolution surface, nahcolite occurs in three principal ways: (type 1) coarse-crystalline aggregates and rosettes and fine-grained nodules scattered through oil shale, (type 2) disseminated, closely to loosely packed crystals of nahcolite in oil shale that commonly form laterally continuous units, and (type 3) beds of either microcrystalline light brown nahcolite, coarse-crystalline white nahcolite, or thin layers of brown microcrystalline nahcolite interbedded with coarse-crystalline halite. About 60% of the resource occurs as type 1 nahcolite, 25% as type 2 nahcolite, and 15% as type 3 nahcolite. The area of nahcolite-bearing oil shale, where it is 100, or more, feet (33 m) thick, occupies about 257 square miles (666 km^2) in the northern part of the Piceance Creek Basin (**Figure 8**). In ten core holes drilled in the central part of the evaporite deposit, the nahcolitic rocks range from about 550 to 1100 feet (168 to 335 m) in thickness. The nahcolite resource in the basin was estimated at 32 billion short tons (29 billion mt) by Dyni (1974a) and 29 billion short tons (26 billion mt) by Beard, Tait, and Smith (1974). Details of the economic geology of the Colorado nahcolite deposits have been reported by Cole and others (1982); Cole and Daub (1983); Cole and others (1995); Nielsen (1995); and others.

In 1970 to 72, Shell Oil Company experimented with warm water and steam injection for recovery of shale oil from nahcolite-bearing oil shale on a 10-acre site in the W T1S, R97W, Rio Blanco County, Colorado. In addition to oil, Ireson (1990) reported that large amounts of sodium carbonate were co-produced with the oil (**Figure 9**).

In 1990, the White River Nahcolite Minerals Company (White River Minerals), a subsidiary of North American Chemical Company, began solution mining the Boies nahcolite bed on a 8222-acre (3327-hectare) Federal sodium lease tract located in T1S, R98W, Rio Blanco County, Colorado (**Figure 8**). The mining method involves drilling a vertical hole to within several hundred feet above the bed, deviating the hole so that it enters the bed at a slight angle, then drilling the hole horizontally along the base of the bed. Hot water injected into the bed dissolves the nahcolite and forms a horizontal cavity. Sodium carbonate brine is recovered from another vertical hole drilled into the cavity some hundreds of feet from the injection hole. Where mined, the Boies bed is about 25 feet (7.6 m) thick and lies at a depth of 2000 feet (610 m) (Day, 1994).

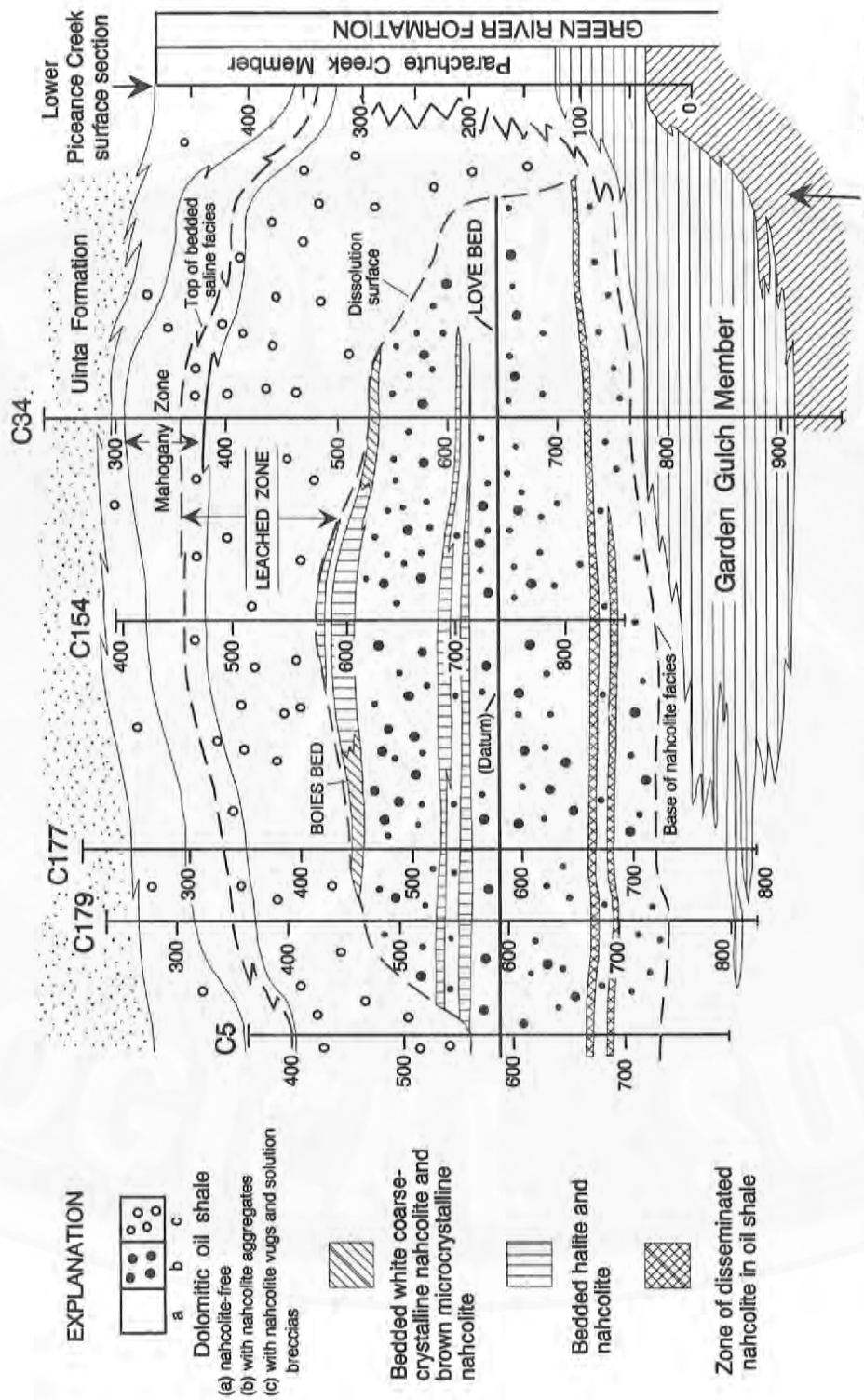


Figure 7. Generalized stratigraphic cross section of the nahcolite deposits in the Parachute Creek Member of the Green River Formation in the northern part of the Piceance Creek Basin, Colorado. Line of cross section is shown in ***Figure 8****. Depths in meters.*

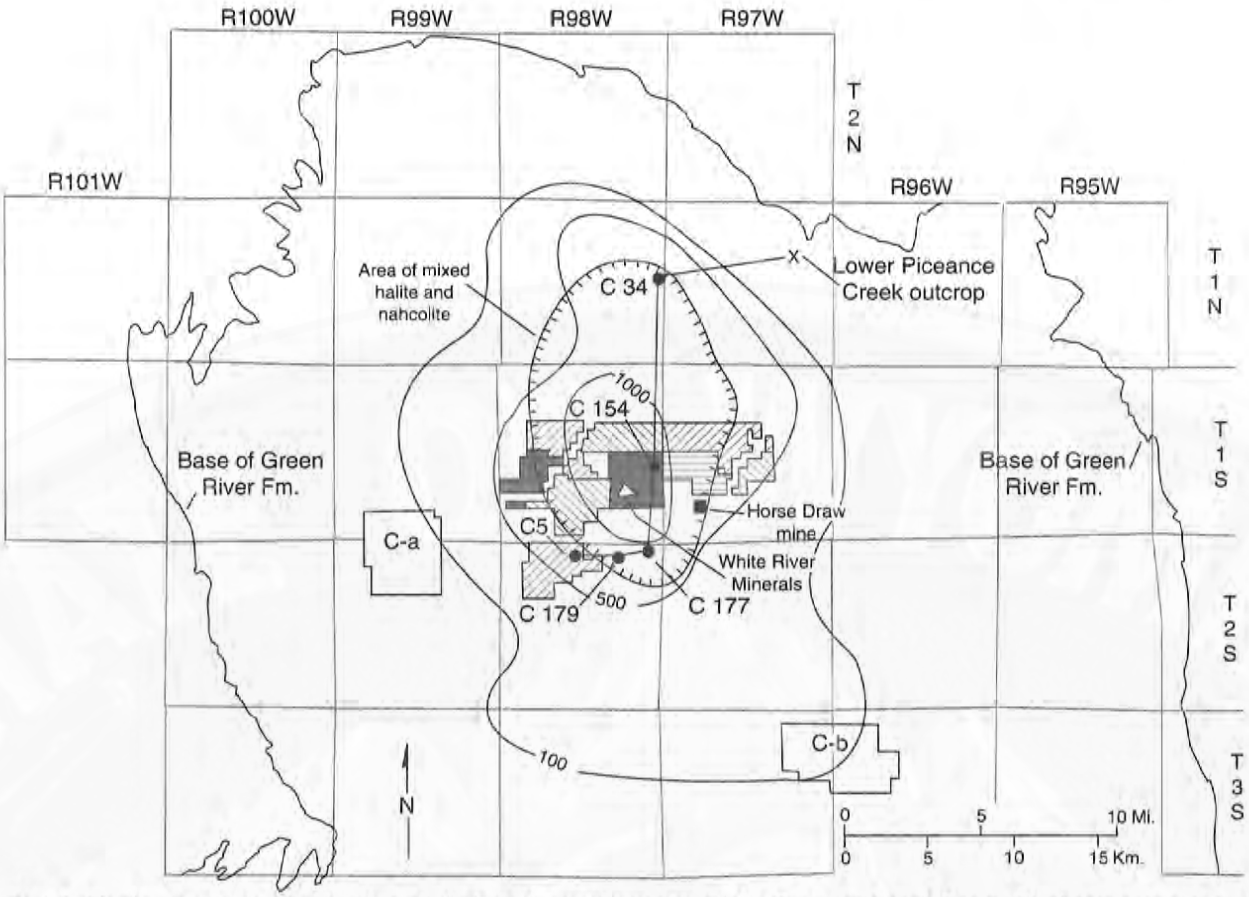


Figure 8. Thickness of nahcolite-bearing oil shale in the northern part of the Piceance Creek Basin. Thickness contours in feet. Patterned areas are active Federal sodium leases. C-a and C-b are inactive Federal oil-shale lease tracts.

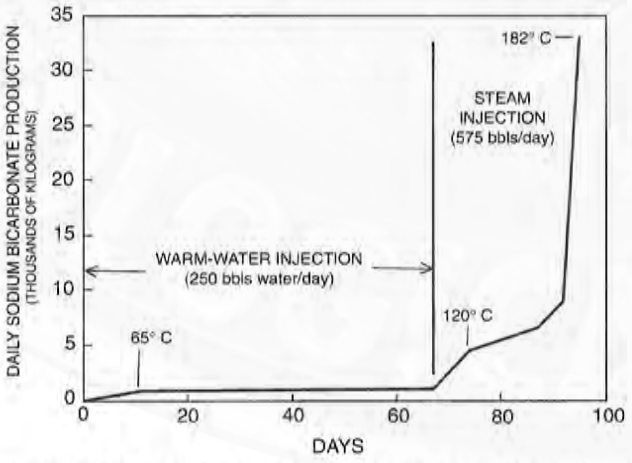


Figure 9. Daily production of sodium bicarbonate from Shell's in situ shale-oil recovery experiment in the Piceance Creek Basin (from Ireson, 1990).

Currently, eight Federal sodium leases have been issued on about 16,900 acres (6839 hectares) in the basin (**Figure 8**). Over the past few years, another company has been drilling exploratory core holes in the basin for possible leasing of sodium minerals.

In addition to the Boies bed (L-5A), several other nahcolite beds in the Piceance Creek Basin may be economical to solution mine including the L-4D (Love bed) and R-5C, R-5E and R-5G beds (Dyni, 1981, plates 1 and 2).

Solution mining nahcolite in the Piceance Creek Basin presents some unusual problems. Because the Boies bed is close to the dissolution surface (**Figure 7**), a breakthrough into the overlying water-bearing leached zone could be detrimental to solution mining in the bed. Furthermore, the Boies bed grades laterally into mixed halite and nahcolite toward the basin's chemical depocenter which could contaminate mining solutions with sodium chloride. The Boies bed as shown by Dyni (1981, plates 1 and 2) actually consists of two beds, the lower one being thinner, but also containing halite. Because oil shale cannot be mined or damaged under current sodium leasing regulations, the major part of the nahcolite resource, which occurs as crystalline aggregates scattered through oil shale, will be difficult to mine for soda ash alone. Solution mining of nahcolite beds in the basin may render parts of the overlying oil shale unsuitable for future *in situ* or underground mining.

Eden-Farson brine field

H.B. Lindeman (1954) first called attention to amber sodium carbonate brine found at depths of 400 to 600 feet (122 to 183 m) in the Green River Basin near the communities of Eden and Farson, about 30 miles (48 km) north of Rock Springs, Wyoming. The discovery of brine was made by the John Hay water well drilled in 1938 in SE sec. 4, T23N, R106W, Sweetwater County, Wyoming. Subsequent exploratory drilling (**Figure 10**) identified many drill holes that encountered black sodium carbonate brine, also known as "black trona water" (Dana and Smith, 1973, 1976). Analyses of brine from four wells by Phillips (1978, table 2) revealed as much as 4 to 6 weight % (wt. %) sodium carbonate-bicarbonate and as much as 5.8 wt. % of water-soluble humic and fulvic acids. Even larger amounts of these components were reported from other wells by Dana and Smith (1973, 1976).

In 1992, a shallow well was drilled for water by the U.S. Bureau of Reclamation in Bone Draw in the NE SE sec. 32, T24N, R107W, about 6 miles (10 km) southwest of Eden. The well encountered an artesian flow of black sodium carbonate brine at 460 feet (140 m). The well flowed an estimated 29,000 barrels (4611 m^3) of brine in two weeks before it could be sealed. A sample of the brine that was analyzed by the Water Resources Division of the U.S. Geological Survey, Denver, Colorado, and the results are given in **Table 2**.

Table 2. Chemical analysis of brine from the U.S. Bureau of Reclamation Bone Draw well 1. Concentrations in milligrams/liter, except where noted; (µg/l) = micrograms/liter.

Dissolved species	Concentration	
Carbonate	13,080	
Bicarbonate	8650	
Chloride	150	
Fluoride	78	
Sulfate	0.8	
Sodium	22,500	
Potassium	58	
Calcium	3	
Magnesium	1	
Lithium	20µg/l	
Dissolved organic carbon	982	
Alkalinity, as $CaCO_3$	50,500	
pH, lab	9.9	
Conductance, lab	49,900	µ/Siemens

The black sodium carbonate brine is found in oil shale of the Wilkins Peak Member of the Green River Formation. The brine appears to be in several discontinuous permeable zones because water flows and concentrations of dissolved solids, (including water-soluble organic matter) are highly variable between wells (**Figure 10**).

Dana and Smith (1976) concluded from examination of drill core from one bore hole that the brine is confined to one or more bedding planes in otherwise impermeable oil shale of the Wilkins Peak Member. These authors noted that a number of the wells in the brine field are artesian and suggested that the brine in the aquifer is pressurized by lithostatic, rather than hydrostatic or gas, pressure. Because of the unusual type of aquifer, the size of the brine field and the composition of the brine cannot be easily determined.

According to Dana and Smith (1976) the stratigraphic interval in which the brine is found lies between 67 and 93 feet (19 and 28 m) below the top of the Wilkins Peak Member. In the trona district to the southwest, this would place the aquifer just above the highest bed (25) of trona in the member.

The origin of the brine cannot be directly linked to dissolution of sodium carbonate salts inasmuch as solution breccias or cavities were not found in the oil shale enclosing the suspected brine aquifer (Dana and Smith, 1976). However, Culbertson (1966, p. B160) suggested that the source of the sodium carbonate in the Eden area was from dissolution of shortite, which is a common salt in the Wilkins Peak Member, rather than from bedded trona. Although salt cavities are absent in the core studied by Dana and Smith (1976), the brine may have formed farther south in the basin and migrated updip along bedding and fracture planes northward toward the Eden area.

The brine has economic potential as sources of soda ash and organic acids (Phillips, 1976). The precipitated organic acids yield as much as 73.3% oil upon destructive distillation. The organic acids may also find specialized uses, such as a soil conditioner (R.J. Hite, personal communication, 1996). Dana and Smith (1976, p. 12) reported a well capable of flowing at a rate of 6800 to 6900 barrels (1081 to 1097 m^3) of brine per day which could yield 62 tons [56 metric tons (mt)] of soda ash and 130 tons (118 mt) of organic acids.

Several companies have examined the brine for possible exploitation, but none to date have pursued

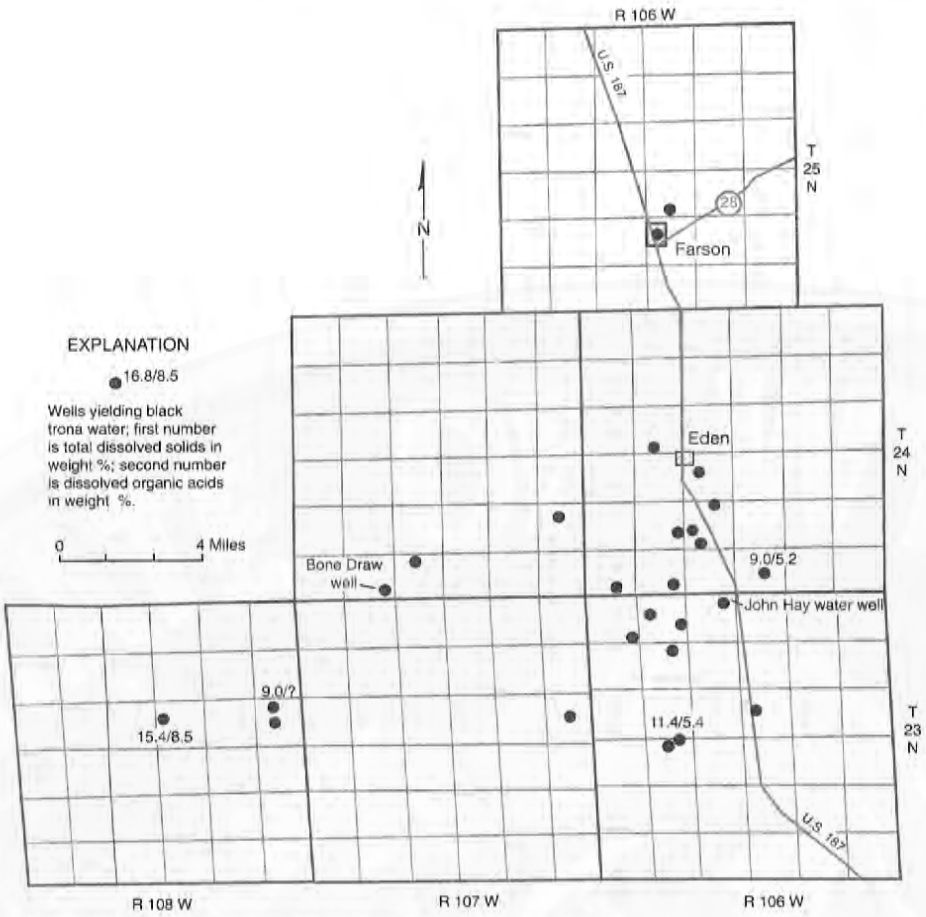


Figure 10. Map showing locations of wells containing sodium carbonate brine and dissolved organic matter in the Eden-Farson brine field, northeastern Green River Basin, Sweetwater County, Wyoming (from Dana and Smith, 1976, and Phillips, 1978).

development of the resource. The Eden-Farson brine field may be producible on a small scale, especially if the organic acids find specialized uses that may command a high unit price. Also, the occurrence of sodium carbonate brine elsewhere in the Green River Basin should not be overlooked.

Wyoming trona deposits

The world's largest known resource of natural sodium carbonate is found in bedded trona in the Green River Formation of the Green River Basin, Sweetwater County, southwestern Wyoming. The formation is divided in ascending order into the Tipton Shale, Wilkins Peak Member, and the Laney Shale. The trona resource lies wholly within the Wilkins Peak Member. The Wilkins Peak ranges from about 600 to 1350 feet (183 to 411 m) in thickness and consists of marlstone, oil shale, trona, siltstone, mudstone, and numerous laterally persistent thin layers of tuff.

As many as 42 beds of trona were noted by Culbertson (1966). He numbered the thickest and economically most important of these beds from 1 to 25 in ascending order (**Figures 11** and **12**). The trona beds are nearly flat-lying. Dips are commonly about 1° or less. Faults are uncommon, however, several strike-slip shear zones have been recently mapped on the surface on the east side of the basin (Gordon Marlatt, University of Wyoming, personal communication, 1996). Some trona beds grade laterally into mixed halite and trona toward the depocenter in the southern part of the basin. In some beds in the lower part of the evaporite sequence, nahcolite is abundant. I have noted from x-ray diffraction studies of mine samples and core samples, that as the depositional center of mixed halite and trona of bed 17 is approached in the southern part of the basin, nahcolite, wegscheiderite, northupite, and possibly shortite become locally abundant, as well as quartz. The occurrence of these additional minerals may be important in room-and-pillar mining because they may influence the processing of trona as well as impart different physical properties to the mined trona.

The generalized lithology of the Wilkins Peak Member in the Union Pacific Railroad, El Paso core hole 44-3, in the southern part of the basin is shown in **Figure 13**. Many of the lower trona beds are shown on the log. Two stratigraphic cross sections showing the distribution of the trona beds in different parts of the basin are shown in **Figures 11** and **12**.

The trona beds are remarkable for their purity. Sulfate minerals are essentially absent, and other than halite, the chief impurities are marlstone stringers and vertical seams of mudstone. Chemical analyses of sev-

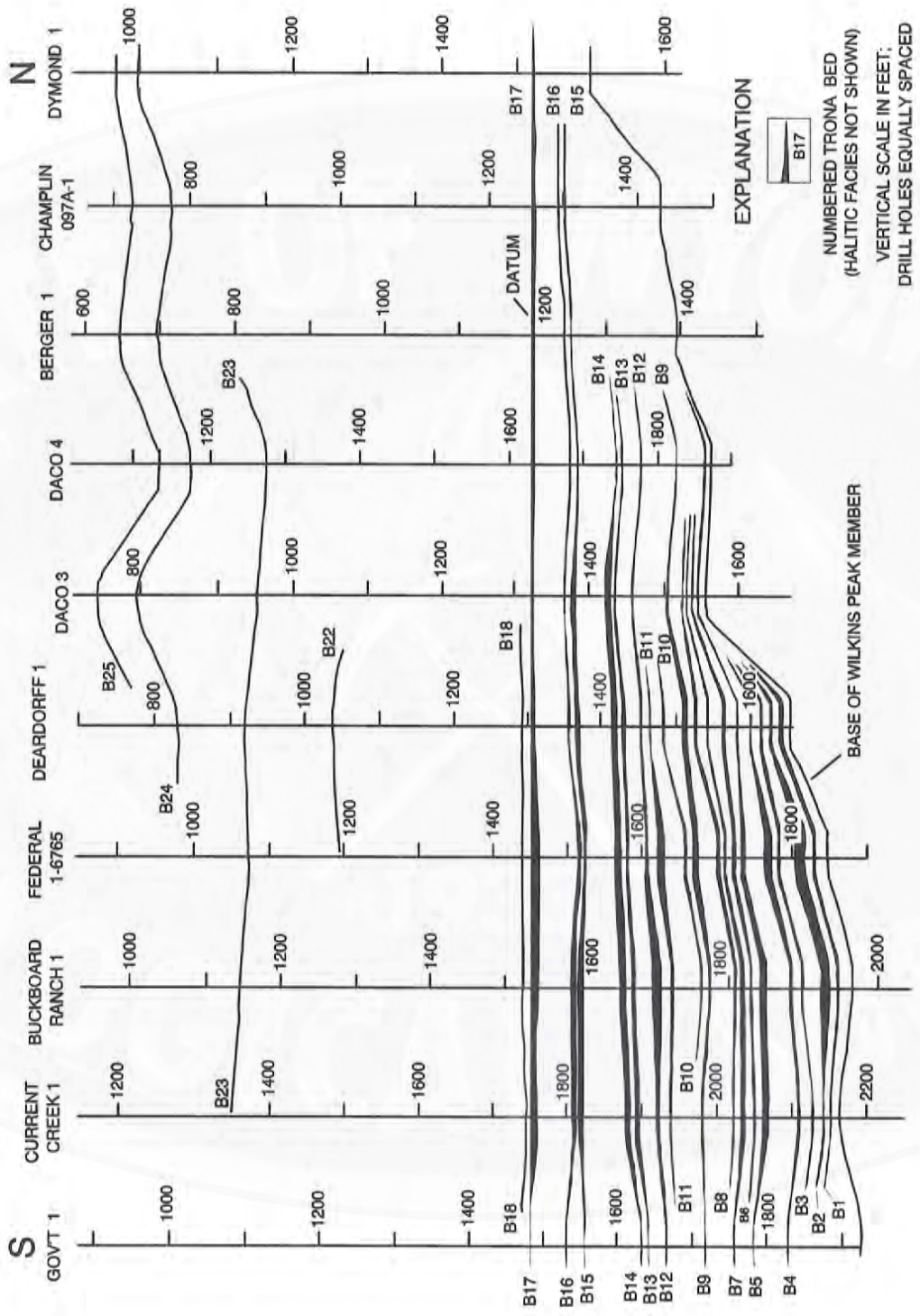


Figure 11. South-north stratigraphic cross section of the trona deposits in the Green River Basin, Wyoming. Line of section shown in **Figure 15**.

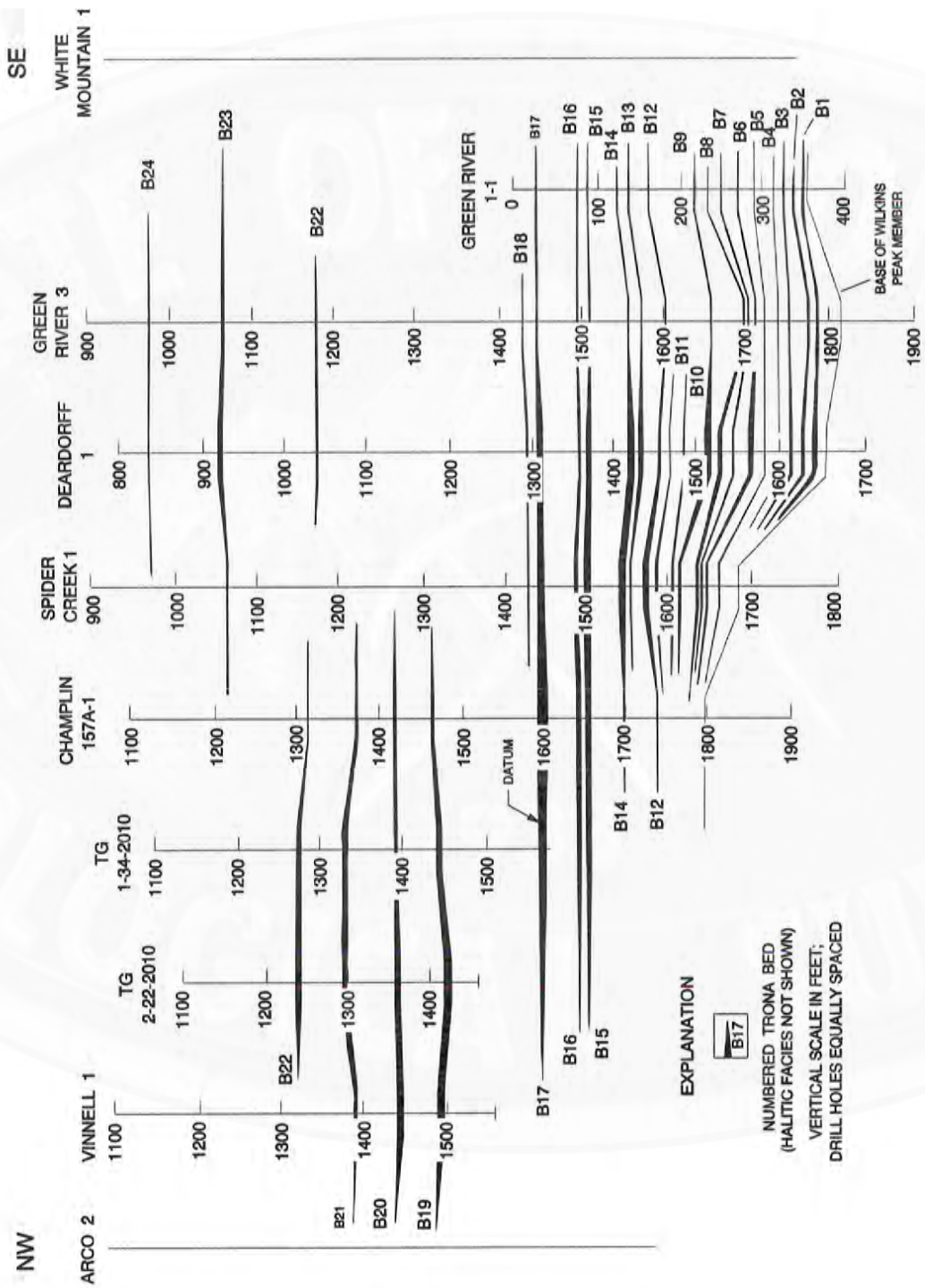


Figure 12. Northwest-southeast stratigraphic cross section of the trona deposits in the Green River Basin, Wyoming. Line of section shown in ***Figure 15****.*

eral trona beds are given by Wiig, Grundy, and Dyni (1995, table 1). Currently, FMC, General Chemical, and Solvay Minerals mine trona bed 17, Tg Soda Ash mines bed 20, and OCi, Inc. mines beds 24 and 25 (Leigh, 1991).

Trona bed 17 is one of the most important trona beds in the district. The bed, where it is more than 4 feet (1.2 m) thick, underlies about 775 square miles (2007 km²) and ranges from 4 to 17 feet (1.2 to 5.2 m) in thickness (**Figure 14**). Southward in the basin, the trona grades laterally into mixed halite and trona.

Twenty-two of the 25 numbered beds are estimated to contain 127 billion tons (115 billion mt) of trona ore, or 76 billion tons (69 billion mt) of trona in beds containing less than 2% halite. These 22 beds, where they are 4 or more feet (1.2 or more meters) thick, underlie areas ranging from 50 to more than 775 square miles (129 to more than 2007 km²) at depths of 650 to 3000 feet (198 to 914 m) below the surface (Wiig, Grundy, and Dyni, 1995). Trona contains 70.3 wt. % sodium carbonate; it takes about 2 tons of trona ore to produce 1 ton of soda ash.

Additional resources of sodium carbonate in the Green River Formation include shortite which occurs in abundant crystals, pods, and veinlets scattered through marly lacustrine strata interbedded with the trona. Less commonly, some shortite may be in bedded layers a few centimeters thick. Shortite contains 34.6% sodium carbonate by weight.

A visual estimate of shortite in drill core from the John Hay core hole 1 in the SE NW sec. 2, T18N, R110W, Sweetwater County, was made by Fahey and Mrose (1962, figure 7). Shortite averaged about 11% for a sequence 550 feet (168 m) thick between the depths of 1250 and 1800 feet (381 and 549 m), and averaged 16.7% for a 230-foot (70-m)-thick sequence between the depths of 1330 and 1560 feet (405 and 475 m). Assuming a similar grade for a one square mile (1.6 km²) block around this bore hole, as much as 203 million short tons (184 million mt) of shortite, or 70 million short tons (64 million mt) of sodium carbonate, per square mile is indicated–a substantial resource.

Because of its mode of occurrence, shortite (or sodium carbonate) would be difficult to mine separately, although some sodium carbonate from the mineral might be extractable by solution mining.

The Wyoming trona district currently produces about 9 million short tons (8.2 million mt) of refined soda ash annually from five underground mines that extract a total of about 18 million short tons (16.3 million mt) of trona ore (Ray Harris, Wyoming State Geological Survey, personal communication, 1996). See papers by Wiig, Grundy, and Dyni (1995), Burnside and Culbertson (1979), and Brown (1995) for additional resource, production, and mining details.

Mining is by vertical shaft using room-and-pillar, short- and long-wall techniques. At three mines, additional sodium carbonate is recovered from processing plant waste waters which are used to slurry tailings to abandoned parts of the mines. These waters dissolve additional trona from the pillars, then are pumped back to the surface to recover the sodium carbonate. The technique has proven to be economical and will probably be employed by the remaining mines in the district.

Earlier attempts to solution mine trona from vertically drilled holes have been unsuccessful. However, solution mining using horizontally drilled holes as used in the Piceance Creek Basin, Colorado, has not yet been attempted; the method will probably be tested soon in the Wyoming trona district. The method may be especially useful in the southern part of the basin where the trona beds are thicker and lie at greater depths than those being mined today.

Solution mining appears to offer major advantages over room-and-pillar techniques. Drilling injection and production wells should cost less than the cost of sinking shafts and opening underground workings. Solution mining will provide safer operating conditions because underground miners will not be needed. Room-and-pillar mining of trona beds in the deeper part of the basin, which would require larger pillars for safety and lower the recovery of trona ore, should pose no serious problems for solution mining. Concurrent solution mining of multiple trona beds should also be feasible.

Approval of mining plans should assure trona-land owners that important trona beds are not made unmineable by multi-bed solution mining. The plans should also assure that land subsidence is controlled, potential oil and gas production is not impaired, and significant parts of the trona resource are not lost by high-grading the thicker and purer beds of trona in the lower part of the deposit.

Ownership of the lands in the trona district is divided between the Federal Government (55.7%), Union Pacific Resources as part of a railroad land grant (38.1%), and the State of Wyoming (6.2%). **Figure 15** shows the boundary of the Federal Known Sodium Leasing Area, and two lines of stratigraphic sections. A good summary of the soda ash industry was published recently by Kostick (1995).

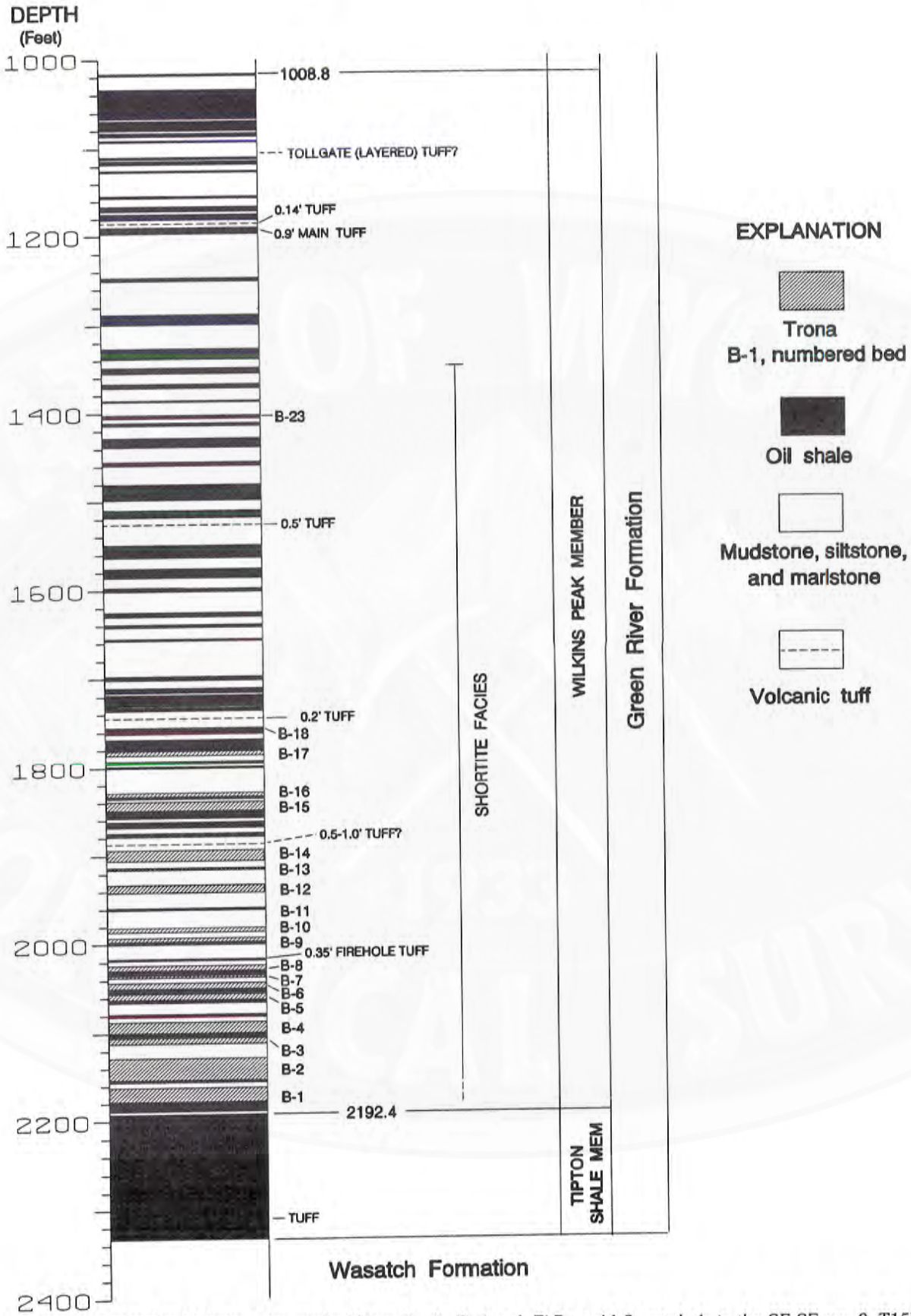


Figure 13. Generalized lithologic log of the Union Pacific Railroad, El Paso 44-3 core hole in the SE SE sec. 3, T15N, R109W, Sweetwater County, Wyoming (modified from Trudell, 1975).

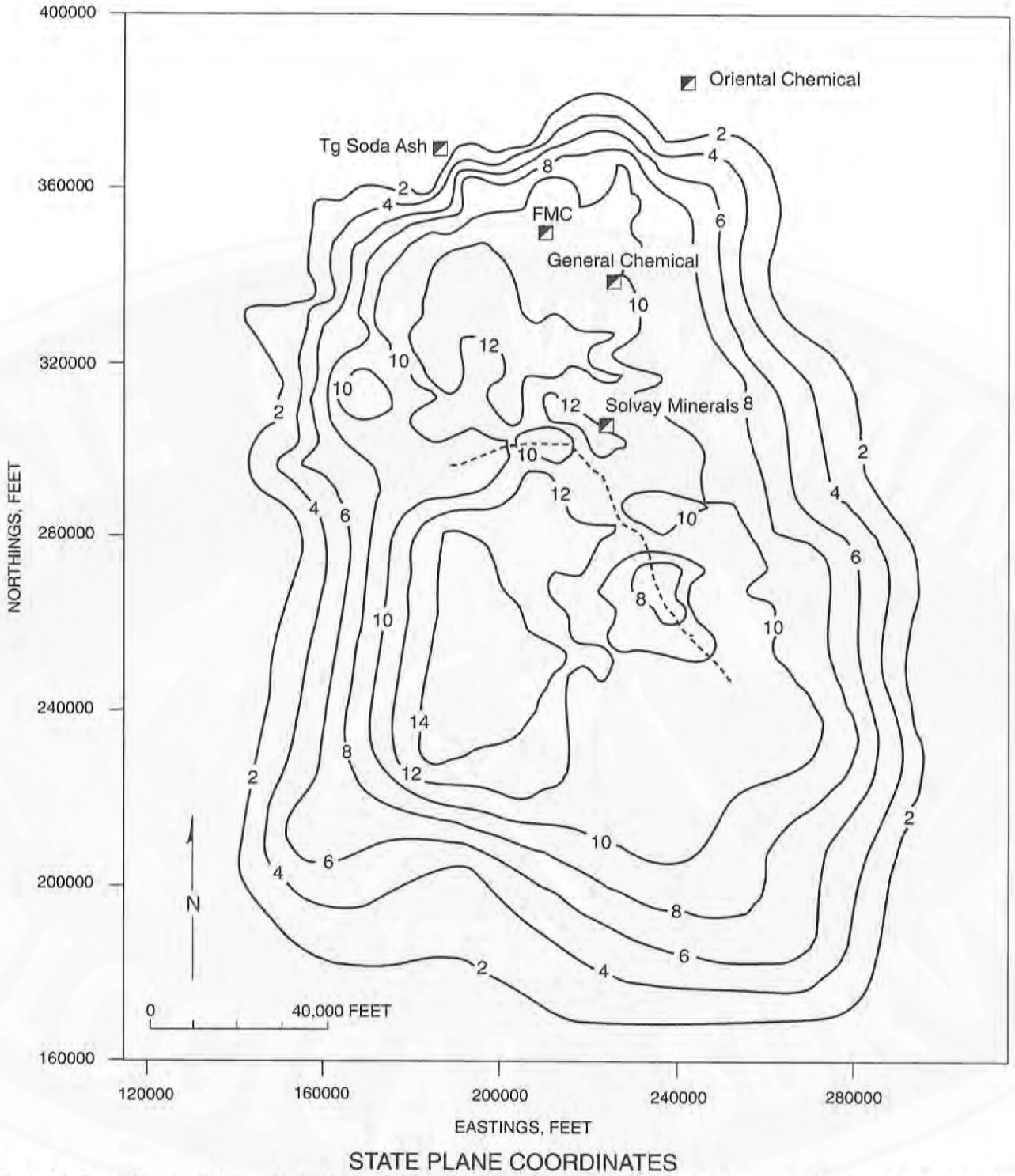


Figure 14. Isopach map, in feet, of trona bed 17. Dotted line indicates a trough of thinned trona in the central part of the bed (modified from Wiig, Grundy, and Dyni, 1995).

The trona resources are summarized in **Figure 16**. Trona bed 2 underlies 439 square miles (1137 km²) where it is 4 or more feet (1.2 or more meters) thick and contains the largest resource of trona of any bed, estimated at 11.6 billion short tons (10.5 billion mt), of which 97% of the bed is essentially halite-free. Trona bed 17, where it is 4 or more feet thick, underlies about 775 square miles (2007 km²) and contains almost as much trona–11.5 billion short tons (10.4 billion mt), but only 66% contains little or no halite (Wiig, Grundy, and Dyni, 1995, table 9). Note that the trona beds 1 through 4, and 19 through 25 are free, or nearly free, of halite.

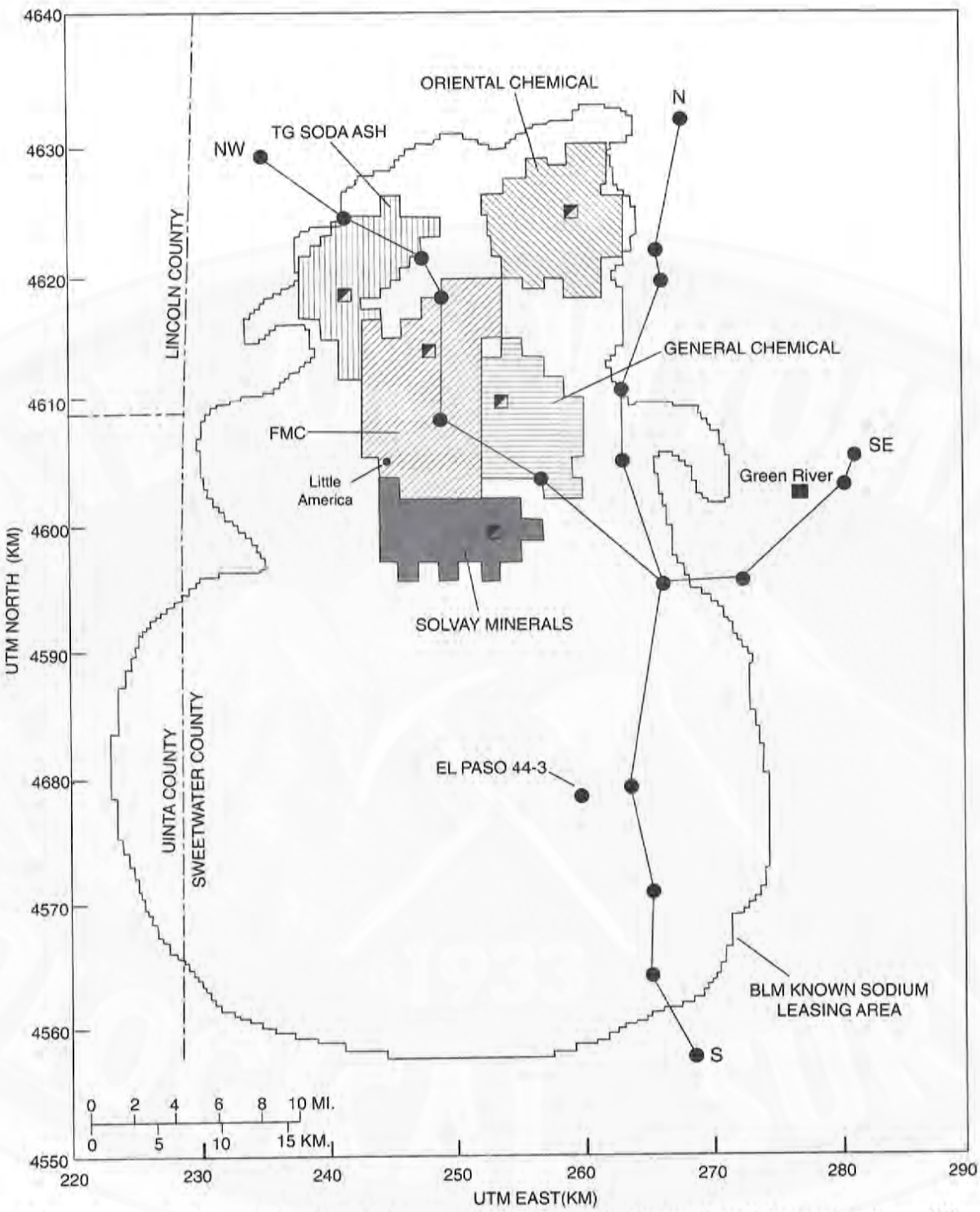


Figure 15. Map showing the Known Sodium Leasing Area established by the U.S. Bureau of Land Management, State of Wyoming trona mine permit areas, and lines of stratigraphic sections shown in ***Figures 11*** *and* ***12****.*

Summary

The geology and economic potential of five occurrences of sodium carbonate mineral and brine deposits in the Eocene Green River Formation were briefly described. The two largest of these are the deposits of bedded trona and disseminated shortite in southwestern Wyoming and bedded and nonbedded nahcolite and fine-grained disseminated dawsonite commingled with high-grade oil shale in northwestern Colorado.

Other lesser known deposits that may have economic potential include bedded eitelite near

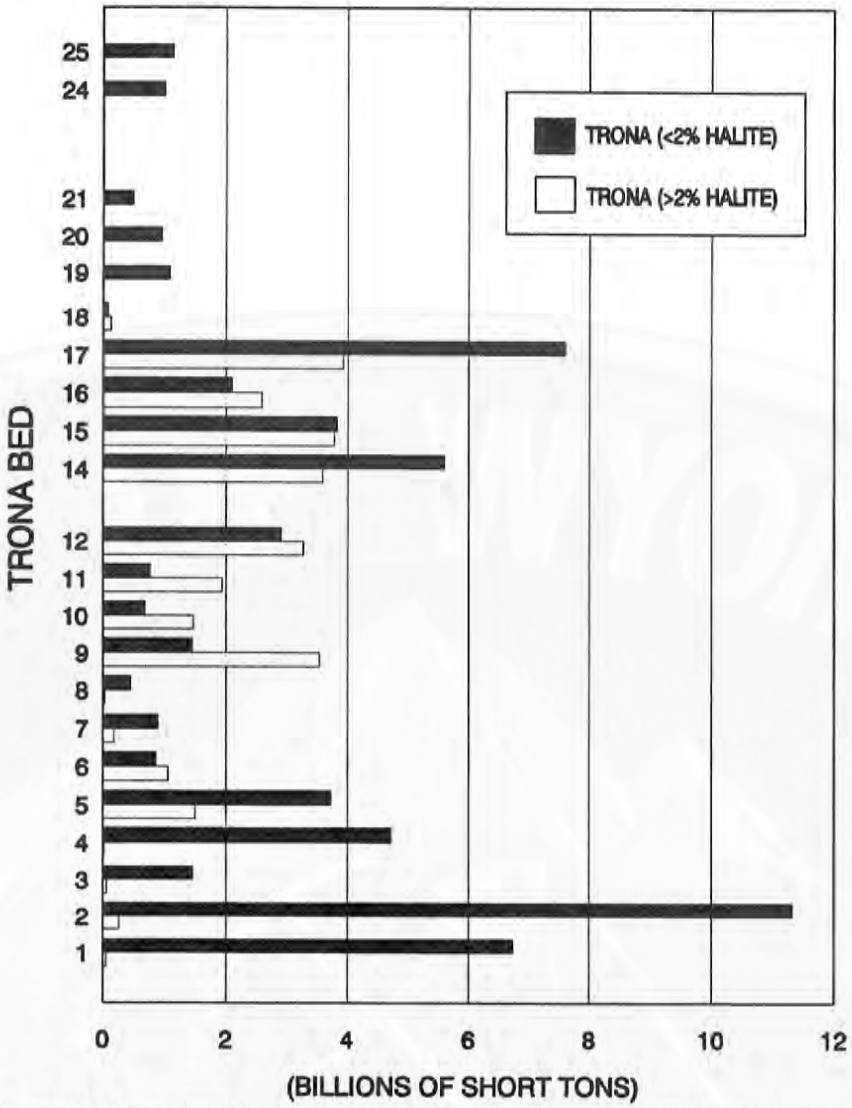


Figure 16. Trona resource estimates, in short tons, for 22 numbered beds of trona in the Wyoming trona district. Data are based on where the beds are 4, or more, feet thick.

Duchesne, Utah and two sodium carbonate brines, one in eastern Uinta Basin, Utah and the other near Farson and Eden in the northeastern part of the Green River Basin, Wyoming.

The sodium carbonate resources in the Green River Formation will continue to provide most of the soda ash to domestic markets and will probably continue to increase their share in foreign markets for the foreseeable future. The technical success of solution mining bedded nacholite in Colorado suggests that a similar operation would be successful in the Wyoming trona district. Solution mining beds of trona through horizontally drilled holes may substantially lower mining costs and improve mine safety. Potential environmental problems such as disposal of plant tailings, waste waters, and stack-gas effluents including carbon dioxide could be reduced by solution mining methods.

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Abandoned epsomite plant north of Rock River, Wyoming. Alkali lakes in the Laramie Basin were the source of the epsomite. The plant was closed in the 1930s. (Photograph by R.E. Harris.)

An overview of gypsum occurrences and resources and the Georgia-Pacific gypsum operations at Lovell, Wyoming

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Abstract

Gypsum is di-hydrated calcium sulfate, $CaSO_4 \cdot 2H_2O$. It often occurs with anhydrite, which is also calcium sulfate but with no water of crystallization. Gypsum is one of the most common naturally occurring sulfate minerals. It is found in very extensive bedded deposits all over the world and is associated with most of the normal-type sedimentary rocks–limestones, shales, sometimes sandstones, marls, and clays.

Gypsum has been used since earliest recorded history, from carvings and decorations to a very poor plaster and mortar as used in the pyramids. In early Greek times the burning of gypsum and the preparation of gypsum plaster was described. This is still the principal use of the mineral today. Gypsum later became known as plaster of Paris because of the well-known deposits in a section of that city.

Gypsum was discovered in New York in 1792, Virginia in 1835, Michigan in 1840, Ohio in 1850, Iowa in 1872, and in Colorado and California in 1875. In 1885, a commercial method of retarding plaster set was developed which revolutionized the industry by permitting the use of plaster in construction. The development of prefabricated wallboard further revolutionized the industry.

Uses of gypsum

Raw or non-calcined gypsum is sold for use in cement, agriculture, or fillers. For cement, it is sold to, or more accurately, purchased by, the cement companies for its SO_3 content which can range anywhere from about 37% for an 80-purity gypsum to about 59% for a 100% anhydrite. The individual cement plants determine what SO_3 content best suits their own needs and then request rock usually within a range of plus or minus 2%. This Portland cement (PC) rock is generally shipped in hopper type rail cars and is sized minus 1 1/2 inches plus 3/8 inches.

In the agricultural field, gypsum is sold as "landplaster" which is rock ground to minus 100 mesh. Gypsum is an amendment or a conditioner and not a fertilizer as such. It is used to aid in water infiltration where high sodium soil conditions exist. Sodium cations exist in soils throughout the U.S. Those soils with excessively high levels of sodium are referred to as sodic soils and are usually found in the arid west and southwest. Irrigation waters can contain high levels of sodium. The highest tonnage gypsum operation in the U.S. used to be in the Lost Hills area of California, northwest of Bakersfield, which produced gypsite, a natural form of landplaster, for the farms in the San Joaquin Valley.

Gypsum separates naturally into calcium and sulfate. In the soil, sulfate forms sulfuric acid which neutralizes any affect the calcium may have on raising pH. Calcium replaces the sodium on the cation exchange sites. The sodium and sulfate then form sodium sulfate which is leached from the soil through water application. Soil particles then come together to improve soil structure and drainage. Gypsum is sometimes used to alleviate the hardness and crustiness of clay soils. This is a temporary improvement as long term success is achieved through addition of organic matter. Gypsum also provides sulfur and catalytic support for maximum fertilizer utilization and leguminous productivity. The

peanut industry in the southeast is a very large consumer of landplaster.

When used as a filler, gypsum is generally very high purity, very white, and very finely ground. It is used in paper, paint, food, toothpaste, and the like. Recently, a television advertisement aired in which a little boy would not eat a brand of ice cream because it contained calcium sulfate. At least it was one of the ingredients that he could pronounce. At one time gypsum was added to animal feed pellets on the theory that it would reduce the smell associated with feedlots. If this is being done, I can assure you that the hog and chicken producers in Iowa are not taking advantage of this opportunity.

Calcined gypsum is used for wallboard or plaster. Because it is very labor intensive, plaster is no longer used in the building industry except for special situations and applications. Plaster is used for roof decks and floor underlayment–in both cases it is pumped on as a slurry, screeded, and allowed to set. Plaster is still used in the casting and molding industries. Wallboard is by far the largest use of gypsum today in the U.S. (**Figure 1**). In thicknesses of 1/2 inch and 5/8 inch it is the primary material used for interior walls and ceilings in the housing market. The old familiar black board sheathing is being replaced by wallboard covered with fiberglass matte rather than asphalt-impregnated paper.



Figure 1. Map of gypsum wallboard production in the United States.

Gypsum resources in the U.S.

In the U.S., resources are centered in five main areas: The Great Lakes area which includes the Silurian Basin and the high quality reserves of the Michigan Basin; the Gulf Coast embayment of south Texas, Louisiana, and Mississippi; the Permian Basin of New Mexico, north Texas, Oklahoma, and Kansas; several basins in the Rocky Mountain area including those of Montana, Wyoming, Colorado, Utah, and New Mexico; and southern California and southern Nevada.

Gypsum deposits may be found in any geologic era but they are most common in the Permian. In the Bighorn Basin, Wyoming, gypsum is found in rocks of Pennsylvanian, Permian, Triassic, and Jurassic age. Near Lovell, Wyoming it is quarried from the Jurassic Gypsum Spring Formation (Figure 2). Some reports refer to this as the Chugwater Formation.

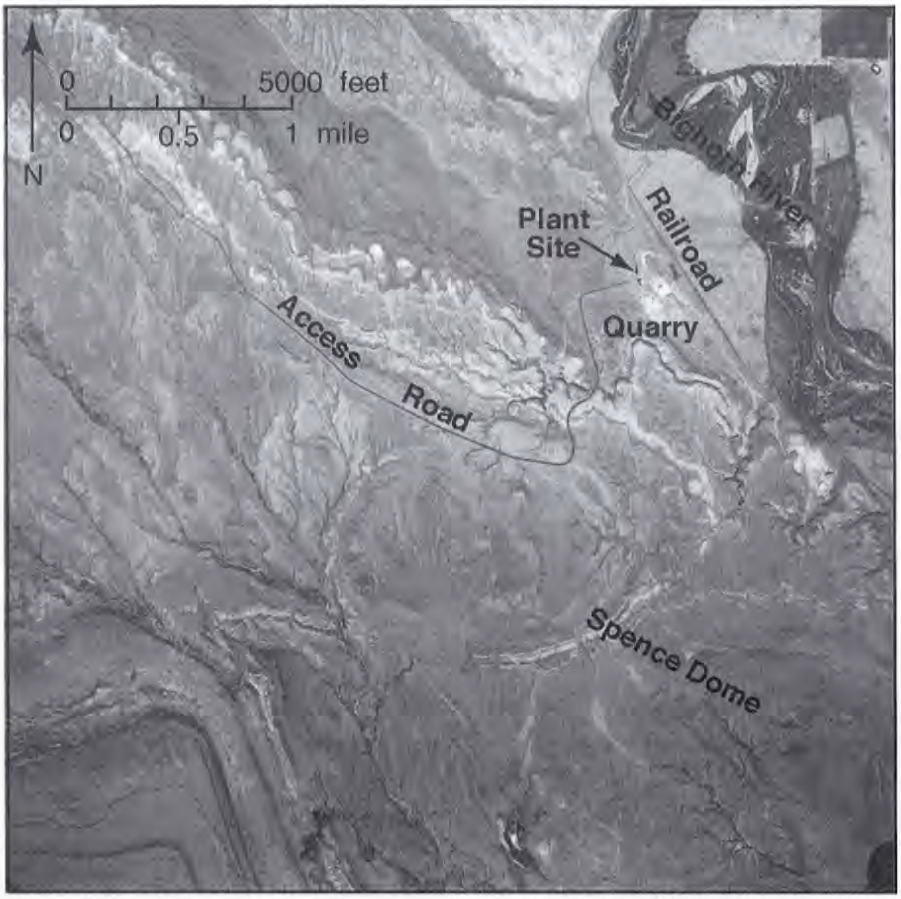


Figure 2. Aerial photograph of the area around Georgia-Pacific's Lovell gypsum quarry and plant.

Formation of gypsum deposits

Calcium sulfate, usually thought to be deposited originally as anhydrite, was probably deposited originally as gypsum in the Bighorn Basin. With burial and compaction, the gypsum normally recrystallizes as anhydrite. With uplift and erosion of overlying sediments, the anhydrite commonly will recrystallize to gypsum. At depth, anhydrite is the stable form of calcium sulfate while at the surface, gypsum is the stable form.

When anhydrite recrystallizes to gypsum it will tend to occupy more space since gypsum incorporates two water molecules in the crystal lattice. The addition of this water of crystallization is the primary cause of the expansion. During the resultant expansion, veins of gypsum commonly form in surrounding sediments or rocks. Some contortion of the original bed and perhaps the surrounding sediment is likely due to gypsum recrystallization.

Some economic aspects of gypsum production

One of the major costs in the gypsum industry is freight–either getting the rock to the plant or getting the product to market. The ideal situation as far as freight costs are concerned would be having a deposit very close to a large metropolitan area. This is not generally the case, especially since there are no gypsum rock deposits on the east coast and essentially no gypsum rock on the west coast. Almost all the raw material used at the plants on the east coast is imported from Nova Scotia; the material used on the west coast comes from San Marcos Island in the Gulf of California. Except for two plants on the Great Lakes, most of the other gypsum plants have a mine or quarry either on site or within a reasonably short distance.

Also because of transportation costs, most plants were built on rail lines so that a side of track could be built into the plant to provide rail cars for loading. Because of competition and service requirements in recent years, wallboard needs to be delivered directly to the job site in the quantity required for the particular job, which requires trucks. Essentially all wallboard is now trucked.

Georgia-Pacific's Lovell, Wyoming operations

At the Georgia-Pacific Lovell quarry (**Figure 3**) the 30- to 50-foot gypsum seam is actually 9 to 11 gypsum veins separated by, or interbedded with, seams of red shale (**Figure 4**). If the shale were any color but red, mining would be much easier. People just think good plaster has to be white rather than red.

Before mining, topsoil is removed and either stored or spread on an area that is already graded and contoured. At the present time overburden is removed by dozers or a combination of cast blasting and dozers. Cast blasting is accomplished by drilling 6" diameter holes at a 15° angle toward the toe in the direction the material is to be moved. The drilling is contracted. After the cast blast the dozers push the remaining material onto the bench. Cast blasting is only done on the deeper overburden. Along the outcrop the shallower overburden is dozed onto the bench. After the overburden is removed the loader cleans out the pockets and removes the remaining loose material in preparation for drilling.

The first four gypsum veins are each mined separately. The top vein (5 to 6 feet thick) is drilled and shot. The shot rock is hauled and crushed. Some rock is screened out for sale as PC rock. Fines from the screen have to be hauled around and worked into the top vein stockpile. After the top vein is removed, the dozers and the loader clean the shale off the second vein (18 to 20 inches thick) which is then drilled and blasted. The rock is hauled, crushed, and screened. The fines are not useable as gypsum so they are hauled back to the quarry. The third vein (30 to 36 inches thick) is cleaned, drilled, and blasted. The rock is hauled, crushed, and screened. These fines are also hauled back to the quarry. About 70% of the second and third gypsum veins is recovered. The fourth vein (5 to 6 feet thick) is quarried the same as the top vein.

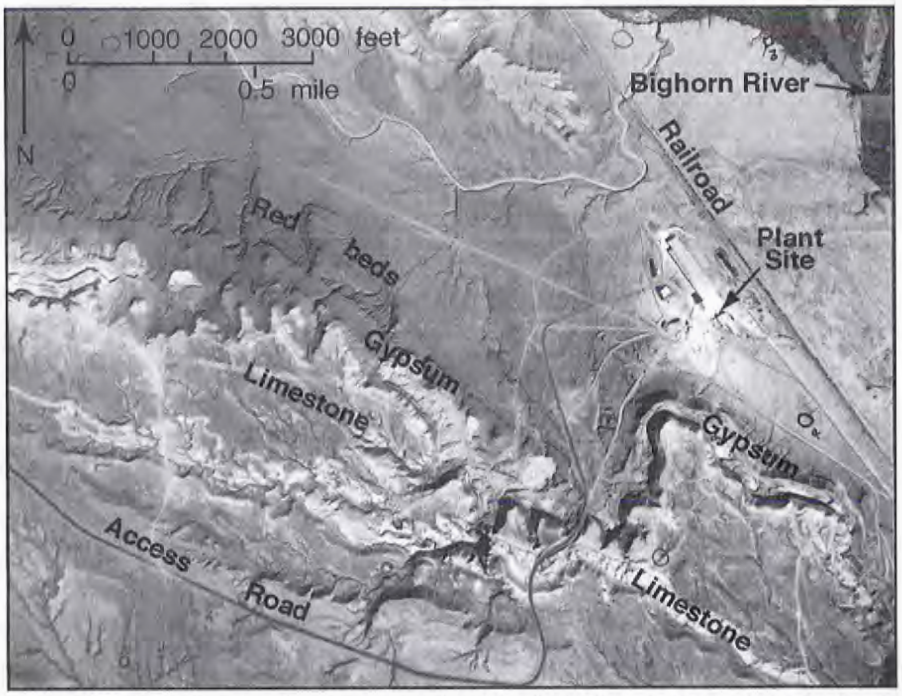


Figure 3. Aerial photograph of Georgia Pacific's plant site, access road, and gypsum outcrop below ridge. Most white color is limestone.



Figure 4. The top 30 feet of the so-called 50-foot gypsum seam.

The fifth through eighth veins are drilled and shot together then crushed and screened with the screening going back to the quarry. About 60% of this gypsum is recovered.

Between the eighth and ninth (bottom) veins are several feet (2 to 6 feet) of waste rock that is ripped (if possible to rip) or drilled and shot (if not possible) and is wasted in the

pit. Then the bottom vein (8 to 12 feet thick) is taken. It is handled the same as the top and fourth veins.

Because of the time involved in stripping, cleaning, drilling, and hauling, three or four areas are worked at the same time, creating a stair-step sequence (**Figure 5**). The quarried areas are back-filled with overburden and screenings. Those areas are graded, then seeded



Figure 5. Mining area at Georgia-Pacific's Lovell gypsum quarry. This shows the stair-step sequence: far end - top vein being worked; next - 2nd and 3rd veins gone–ready for 4th; next - 4th vein gone - 5th-8th veins drilled; next - 5th-8th veins gone–shale pushed off; and finally, bottom vein drilled.

(**Figure 6**). Sprouting success depends on the weather. At one time most of the areas that had been quarried had been reclaimed. Some of the land has been redisturbed as more of the material downdip under the thicker overburden is quarried. Presence of anhydrite and/or sodium sulfate impurities which increase downdip will determine the mining limit.

The plant (**Figure 7**) recycles over 100% of the waste produced. All of the imperfect material that is produced from start-ups, shut-downs, paper breaks, etc., is put back through the system the next day. In addition, the plant has been opening old dumps, removing the wallboard, processing it through a tub grinder or hay buster, and then passing it back into the system to make new wallboard.



Figure 6. Graded and reclaimed quarry area. This shows the bench after final grading. Area in foreground has been seeded. Color version of this photograph shows a slight tint of green in the area.



Figure 7. Georgia-Pacific's gypsum processing plant near Lovell. A quarry truck is dumping ore into a crusher; the crushed ore then passes through a screening plant and then is stockpiled with conveyers. The truck will then get a load of screenings from a tank and haul it back to the quarry for back filling.

STATE OF WYOMING
1933
GEOLOGICAL SURVEY

Clinker – Its occurrence, uses, and effects on coal mining in the Powder River Basin

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Abstract

Clinker–rock that has been baked or melted by the burning of underlying coal beds–covers about 1600 square miles (mi²) of the Powder River Basin in northeastern Wyoming and southeastern Montana. The Wyoming part of the basin contains about 500 mi² of clinker–half in the Rochelle Hills east of Gillette and Wright; the remainder lies east of Sheridan and Buffalo, and in the Tongue and Powder River valleys. The Montana part of the basin contains about 1100 mi² of clinker. The clinker in place was produced by the natural burning of some 30 to 40 billion tons of coal. Much additional clinker has been eroded away. Zircon fission-track ages show that most of this clinker has formed within the past three million years. Within historical times most major coal-bed fires have been extinguished, but some are still burning, and new ones ignite from range fires and spontaneous combustion.

As a coal bed burns into a hillside, overlying strata harden and collapse into the void where the coal used to be. The resulting clinker is more resistant to erosion than the unbaked rocks or sediments because it is harder, and being highly fractured, allows water to infiltrate rapidly, thereby reducing runoff. As a result, clinker caps ridges and escarpments and controls the topography of much of the basin. The overall reddish color of clinker is due to the oxidation of iron; darker "chimneys" of harder rock represent areas of intense heating. Up close, the color and texture of clinker vary depending on the original lithology, degree of heating, and degree of oxidation or reduction. Colors range from red, maroon, yellow, and pink in oxidized rocks to gray, black, and green in limited reduced areas. Textures range from the original shale or sandstone to porcellanite, paralava, and even glass.

The interface between coal and clinker along the "burn line" poses four major challenges for coal mining: 1) determining the location of the coal; 2) preventing water inflow problems; 3) maintaining adequate coal quality to satisfy contracts; and 4) dealing with increased mining difficulty. Spontaneous combustion of exposed coal is common in most mines. Drilling, magnetometer surveys, and seismic studies can all be used to delineate the limit of the clinker.

Prehistoric Native Americans used porcellanite for spear points and arrowheads. Today, the main use of clinker is for road surfacing in coal mines and in areas where better quality gravel is not available. It is lightweight and easily excavated but varies in hardness and breaks down easily. Clinker works well as a cement-treated road base and as a chip seal on asphalt roads. Clinker is also used for landscaping, street sanding, and erosion control. Several hundred thousand cubic yards of clinker are sold in Wyoming each year by the U.S. Bureau of Land Management and the U.S. Forest Service.

Introduction

Much of the landscape of the Powder River Basin (PRB) in northeastern Wyoming and southeastern Montana is defined by flat-topped, steep-sided hills and escarpments capped by a layer of reddish rock. This layer is clinker–rock that has been baked or melted by the burning of underlying coal beds, much as bricks are altered by firing. Through geologic time, erosion has lowered the level of the whole landscape in the PRB. Clinker is more resistant to erosion than the unbaked rocks and sediments because it is harder and, being highly fractured, allows water to infiltrate rapidly, which reduces runoff and surface erosion (Coates, 1991). Consequently, clinker remains to form the high ground.

Clinker-capped hills cover many square miles, and the thoughtful observer asks–when did all this coal burn? Coal beds have burned in this region episodically over the past several million years–not as a continuous single fire in any one place, but as many intermittent fires separated in place and time. The subbituminous coal in the PRB ignites easily because of its low self-heating temperature (as low as 86°F), above which oxidation and heating accelerate until ignition occurs (Kuchta and others, 1980). A fire started in the outcrop of a coal bed by spontaneous combustion, lightning, a range fire, or more recently by humans, spreads along the outcrop and then progressively inward from the outcrop beneath overlying beds of rock. Hot gases, carried upward from the burning coal, roast the overlying rock to produce clinker (**Figure 1**). At any given time in the past a great many fires were burning in different places; they ranged from fires inches or feet in diameter that produced little heat or smoke to roaring underground infernos that melted parts of the overlying rocks.

Early explorers, including Lewis and Clark, found abundant coal fires and clinker in the northern Great Plains. Meriwether Lewis, while traveling in the badlands of North Dakota along the Missouri River, was the first to speculate in writing on the relation between the two phenomena. His journal entry (Thwaites, 1969, v. 1, p. 315) of April 16, 1805, states:

> *I believe it to be the stratas of coal seen in those hills which causes the fire and burnt appearances frequently met with in this quarter. Where those burnt appearances are to be seen in the face of the river bluffs, the coal is seldom seen, and when you meet with it in the neighborhood of the stratas of burnt earth, the coal appears to be precisely at the same height, and is nearly of the same thickness, together with the sand and a sulphurous substance which usually accompanies it.*

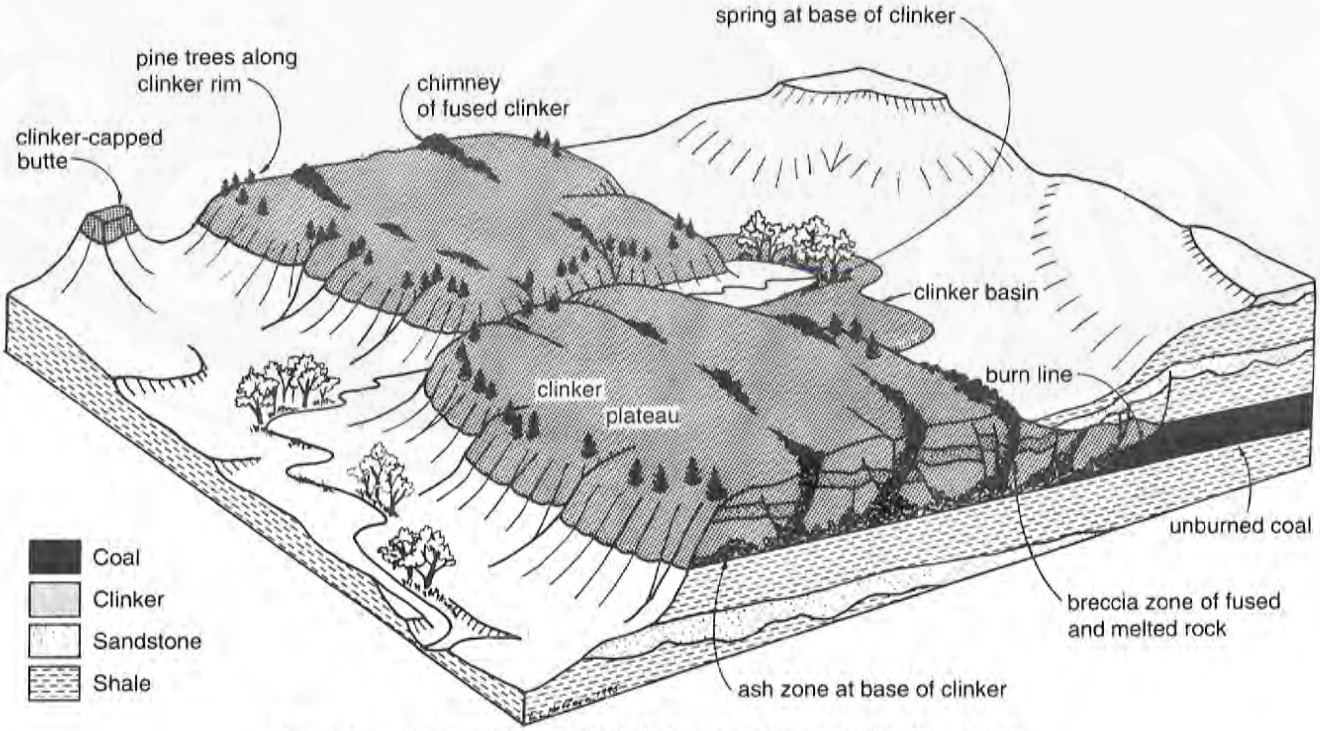


Figure 1. Block diagram of typical clinker landscape, Powder River Basin.

Clark's return route took him along the Yellowstone River, across the northern edge of the basin in Montana. Observing the large amount of red clinker in the gravel at the mouth of the river now known as the Powder River, he named it the Redstone River, and noted that the Indian name for this river, *Wa ha sah*, also meant red stone (Thwaites, 1969, v. 5, p. 310).

Since the late nineteenth century, homesteaders, and later Federal agencies including the Bureau of Land Management (BLM) and the Bureau of Mines, have systematically suppressed most coal bed fires. Consequently, although clinker forms a dominant element in the landscape, we now rarely see the type of intense burning that produced it.

Natural coal bed burning and production of clinker

The term "clinker" encompasses all the rocks that have been altered by heating from a coal bed fire. The term was coined by Prince Alexander Maximilian in 1833, who noted that the rock on the red hills resembled the refuse of European brick kilns, and when struck, emitted a clear sound, "like that of the best Dutch clinkers" (Sigsby, 1966, p. 6). Other commonly used synonyms for clinker are "scoria," "baked and fused shale," and "red dog."

As a coal bed burns, most of the heat liberated is transferred upward by convection (Rogers, 1918) and the strata below the coal remain largely unaltered. The overlying rocks or sediments collapse into the void left by the burning of the coal bed, creating extensive fractures through which hot combustion gases rise. Gases rising along major fissures formed during block slumping ignite and burn with enough intensity to produce darker "chimneys" of harder rock. The interface between the clinker and the unbaked overburden and coal in the subsurface is called the "burn line" or "burn front" (**Figures 2, 3,** and **4**). The term "burn front" may be used to refer either to the front of an active fire, or to the maximum extent of a former fire. The term "burn line" generally refers to an extinguished burn front. This front may be transitional over a few hundred feet, with the coal gradually thinning toward the outcrop across an area where only the top part of the coal bed has burned. The coal adjacent to this boundary is usually oxidized. Clinker is commonly found at depth by drilling as far as several hundred feet back from where it is apparent at the surface. The overlying strata in this zone commonly sag toward the area of completely burned coal. The burn front advances into the hillside until, with increasing depth, fissures in the overburden fail to reach the surface. At that point, the supply of air is cut off, extinguishing the fire. **Figure 4** illustrates typical features of a burn front on a hillside. Where a partial coal seam is present below the clinker, burning may have occurred only above the water table in the seam (Budai and Cummings, 1984).

These processes produce a wide variety of thermally metamorphosed rocks. Most clinker can be described in terms of three variables: the grain size and mineralogy of the parent sedimentary rock; the degree to which heating has affected the mineralogy and bulk characteristics of the rock; and the degree of oxidation or reduction during and after heating, which affects color.

During heating, the changes are progressive: the rock is hardened and reddened in increasing degrees, and fracturing develops. One of the first changes observed is increasing fissility of mudstones, due to

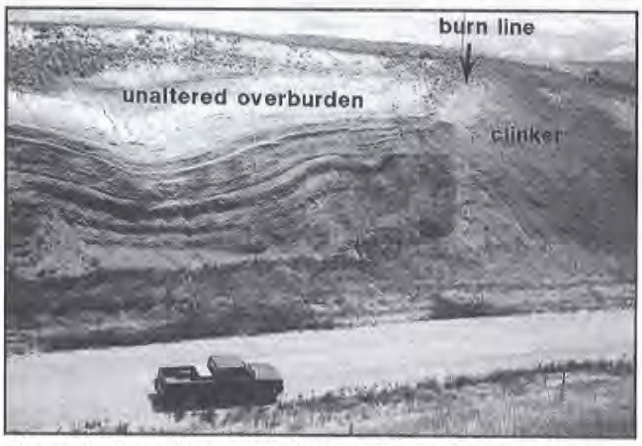


Figure 2. Burn line in roadcut, south side Lake De Smet, Johnson County, Wyoming.



Figure 3. Burn line of Wyodak coal bed exposed in highwall of east pit, Black Thunder coal mine, Wyoming, September, 1993. Note dip of overburden towards area of completely burned coal at right of photograph.

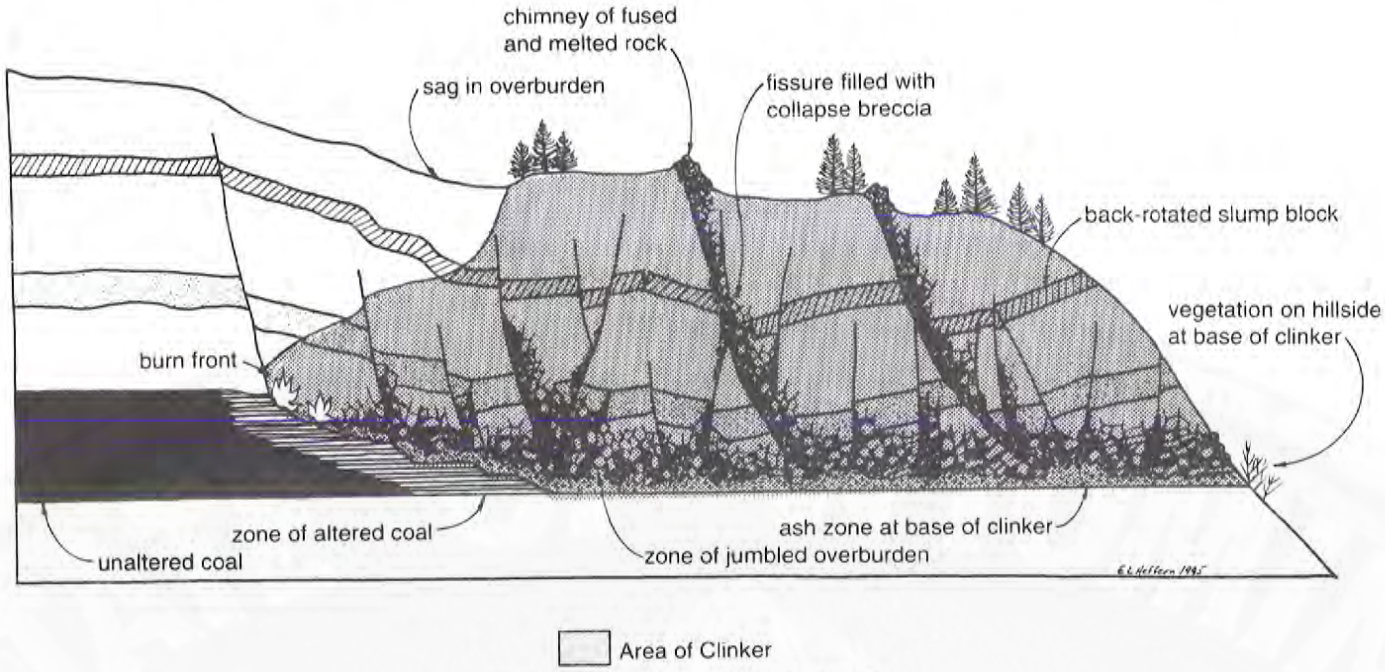


Figure 4. Cross section of typical burn front.

delaminization of phyllosilicate minerals by dehydroxylation (Clark and Peacor, 1992). As heating continues and the rocks dehydrate and oxidize, the shale and siltstone become intensely fractured, and yellow hues give way to darker, more intense shades of orange and red due to the formation of hematite. With further heating, the shale and siltstone recrystallize and take on a fused appearance. Intensely heated shale and siltstone become sintered and the rock develops a ceramic texture, forming porcellanite–a very fine-grained mixture of plagioclase, gypsum, quartz, carbonates, and silica glass (Sigsby, 1966, p. 133). Porcellanite is commonly in a gray reduced form in the interior of joint blocks and in a red oxidized form on joint surfaces. Sandstone changes less during heating and retains its identity until partial fusion occurs (Sigsby, 1966, p. 122). It may take on a reddish color due to oxidation of iron, or even a yellow tint if rich in calcium carbonate.

Responding to variations in temperature and availability of oxygen, some parts of the rock are more intensely oxidized and hardened than others. Generally, the more intense alteration is deeper, nearer the burning coal, and along fissures, or "chimneys," in the overlying rock. Where coal is burning or being intensely heated, gases driven off the heated coal escape into fissures and rise by convection. In these fissures the air supply is generally restricted. When the ascending hot gases encounter sufficient oxygen, they ignite with the intensity and manner of a blast furnace. The resulting incomplete combustion produces carbon monoxide and hydrogen (Sigsby, 1966, p. 98), which reduce the adjacent sediments. Masses of black to green to dark red sintered and melted rocks then form, and cut upward as dike-like chimneys (**Figure 5**) through the less altered clinker. The melted phase of the clinker, called paralava, is formed primarily in the chimneys in an environment of high temperatures (>1300°C) and pressure of only one atmosphere (Cosca and others, 1989). Being of harder rock, these chimneys later stand as irregular masses or walls above



Figure 5. Clinker chimney in Custer National Forest east of Ashland, Montana.

the general surface on clinker areas where less resistant rocks have been eroded away (**Figures 1** and **4**).

A layer of light gray to tan coal ash separates the overlying clinker from the unaltered rocks below. The ash is rarely well exposed except in landslide scars and artificial cuts, or on rapidly eroding slopes. Many fragments of overlying beds that have collapsed into the burning coal bed are churned into the upper part of the ash. In some places intense heating has fused the ash and the base of the overburden, producing a rind of green glass on the surfaces of fragments. Thickness of the ash layer is highly variable, but a few inches to three feet (10 to 100 cm) is a common range in the PRB, where coal beds may be greater than 60 feet (20 m) thick. Some areas have little or no ash possibly because groundwater removed the soluble components in the ash.

Extent of clinker

Exposures of clinker cover about 1600 mi² (4100 km²) of the PRB. In the Wyoming part of the PRB, clinker covers about 460 mi² (1200 km²). Clinker produced by coal beds in the uppermost part of the Paleocene Fort Union Formation covers about 290 mi² (750 km²), while clinker formed by coal in the Eocene Wasatch Formation covers about 170 mi² (440 km²). These figures may be slightly low because clinker commonly is present at depth beyond where the burn line intersects the surface. In addition, some clinker is buried beneath younger alluvium and colluvium. The stratigraphic column in **Figure 6** presents a general view of the main coal beds that form clinker in the Wyoming part of the PRB.

About 1000 mi² (2600 km²) of clinker are exposed in the Montana part of the PRB east of the Crow Indian Reservation (Heffern and others, 1993). About 990 mi² (2560 km²) of clinker are in the upper Fort Union Formation (Tongue River Member) and 10 mi² (30 km²) are in the Wasatch Formation. The largest areas of clinker cap plateaus dividing the Tongue River from Rosebud Creek to the west and from Otter Creek to the east. Extensive areas of clinker also occur west of the Tongue River north of Decker and along the Tongue River valley near Ashland. Crow tribal and ceded lands probably contain about a hundred additional square miles (260 km²) of clinker (Mapel and Griffith, 1981; Robinson and Van Gosen, 1985, 1986).

An average clinker thickness of 60 feet (20 m) was used to make a conservative estimate of the volume of clinker in place (thicknesses as great as 200 feet [60 m] are shown in some mine plan cross sections). Multiplying the thickness of 60 feet by the 1600 mi² (4100 km²) of clinker exposed in the PRB gives an estimate of about 100 billion yd³ (80 billion m³) of clinker in place today. If one further assumes that the 60 feet (20 m) of clinker results from burning of a coal bed 20 feet (6 m) thick–a 3:1 ratio of clinker to coal thickness–then the clinker in place represents 30 to 40 billion tons (33 to 44 billion metric tons) of naturally burned coal. As coal beds up to 200 feet (60 m) thick have formed clinker, this is considered a conservative estimate of coal tonnage. In addition, an undetermined amount of coal, perhaps an order of magnitude greater, burned and formed clinker that was subsequently eroded.

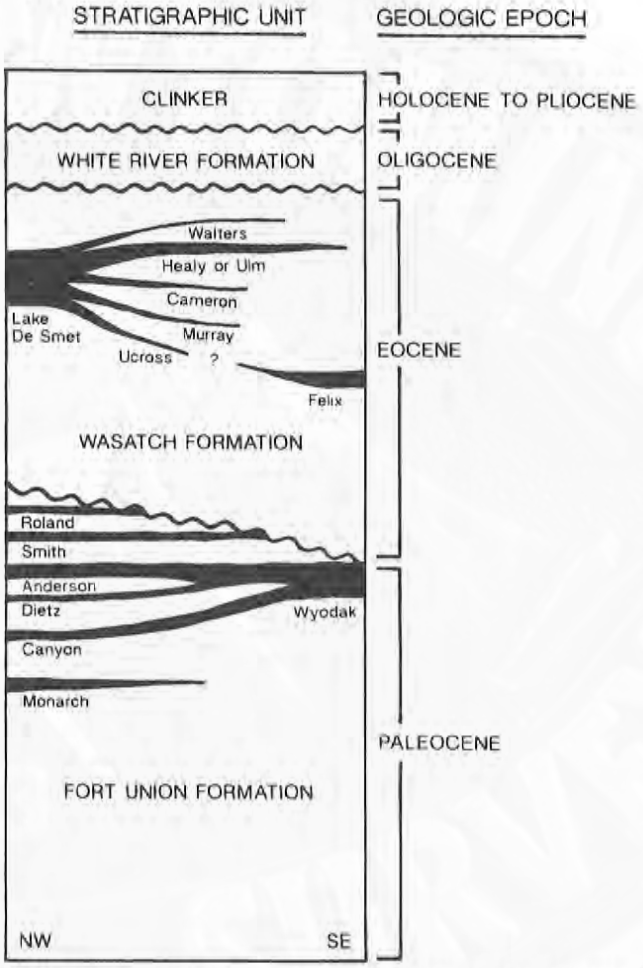


Figure 6. Generalized stratigraphic column showing major clinker-forming coal beds in the Powder River Basin, Wyoming.

Clinker areas within the Powder River Basin, Wyoming

Although the properties of clinker are similar throughout the PRB, this report will concentrate on clinker in Wyoming. Clinker outcrops there can be grouped into five main areas: Rochelle Hills, Healy-Lake De Smet area, Felix area, Powder River breaks, and Tongue River breaks, with widely scattered areas of clinker elsewhere (**Figure 7**). There is very little clinker in the southern third of the PRB in Wyoming, in the area south of the Red Hills (T40N, R70-71W), and in the Miller Hills (T38-39N, R67-68W) in northern Converse County.

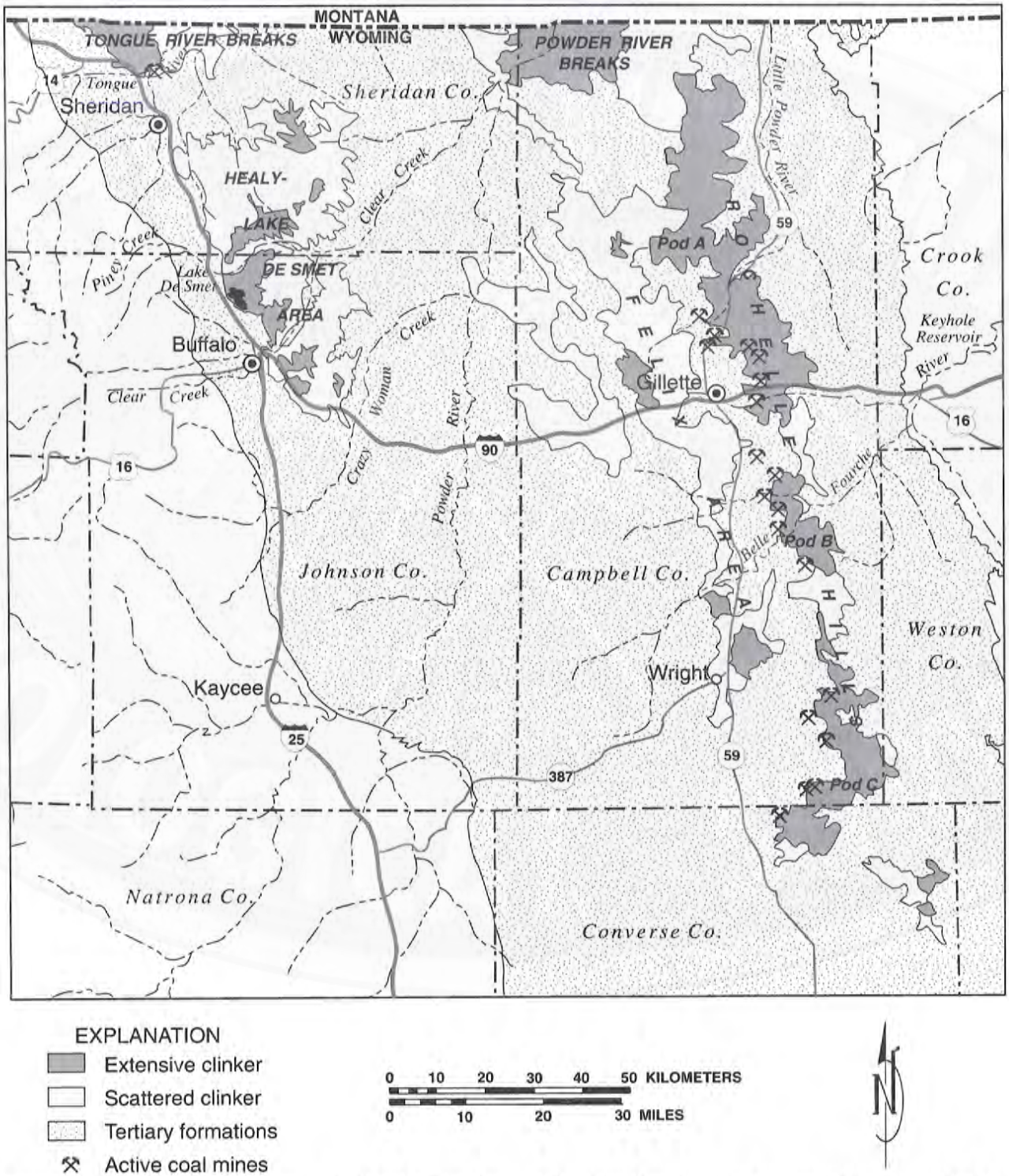


Figure 7. Location of clinkered areas in the Powder River Basin of Wyoming (based on a color map originally compiled by Rob Thurston, Botany Department, University of Wyoming, from the authors' data).

Rochelle Hills

The largest area of clinker, about 230 mi² (600 km²), occurs in a north-south trend along the eastern flank of the PRB east of State Highway 59 in Campbell County and the northern few miles of Converse County. This east-facing escarpment, called the Rochelle Hills (**Figure 8**), is created by clinker formed by burning of the gently west-dipping Wyodak-Anderson and related coal beds in the uppermost Fort Union Formation (**Figure 6**). The thickest clinker is in three main "pods" covering 210 mi² (540 km²) along the escarpment. The areas of thickest surface-minable coal, and all but one of the active coal mines in the Wyoming part of the basin, lie immediately west of and downdip from these three pods (**Figure 7**). The creeks in all the drainages listed below originate above and west of the clinker, then flow east or northeast through valleys cut into the clinker escarpment, with the exception of Piney Creek and one branch of the Little Powder River, which both originate in the clinker.



Figure 8. View to northeast of clinker capping Rochelle Hills escarpment, east of Wright, Wyoming.

Pod A

This pod is located north and east of Gillette, in the drainages of the Little Powder River, and Wildcat, Horse, White Tail, Elk, and Donkey Creeks. The Rawhide, Eagle Butte, Dry Fork, Wyodak, Buckskin, Clovis Point, and Fort Union coal mines lie just west of the southern part of this pod, where coal up to 120 feet (37 m) thick is mined. This pod contains about 130 mi² (340 km²) of clinker (Reheis, 1987; Reheis and Williams, 1984).

Pod B

Located south and east of Gillette, along the drainages of the Belle Fourche River and Caballo Creek (Reheis, 1987), this pod contains about 20 mi² (50 km²) of clinker. The Caballo, Caballo Rojo, Cordero, Belle Ayr, and Coal Creek mines lie just west of the clinker.

Pod C

This pod is located east and south of Wright to just south of the Campbell-Converse county line, in the drainages of Black Thunder, Little Thunder, Piney, Antelope, and Porcupine Creeks (**Figure 9**). Black Thunder, Jacobs Ranch, North Rochelle, Rochelle, North Antelope, and Antelope mines lie west of this pod, which contains about 60 mi² (160 km²) of clinker (Reheis and Coates, 1987).

Healy-Lake De Smet area

The second largest area of clinker, which we have termed the Healy-Lake De Smet area, covers about 110 mi² (280 km²) on the drainages and divides of Piney, Boxelder, Clear, and Dry Creeks and along the divide between Piney Creek and Dutch Creek, east of Interstate 90, north and east of Buffalo (**Figure 7**). This clinker was formed by burning of the Lake De Smet coal zone (**Figure 6**) in the Wasatch Formation (Obernyer, 1978), several hundred feet stratigraphically above the projected position of the Wyodak-Anderson coal zone, which formed the Rochelle Hills clinker. Immediately east of the Bighorn Mountains, the Lake De Smet coal bed is the thickest known coal bed in the United States. The depression that contains Lake De Smet itself (**Figure 10**) was formed by the natural burning of this coal bed, which is up to 200 feet (60 m) thick in the area, causing subsidence of the land surface (Mapel, 1959).

North, northeast, and southeast of Lake De Smet, the Lake De Smet bed splits into several beds, including the Walters, Healy, Cameron, Murray, and Ucross coals (Mapel, 1959; Obernyer, 1978). Accordingly, the distribution of the clinker changes from a single thick pod near Lake De Smet to a series of plateaus, knobs, ridges, and benches to the north and east (**Figure 11**).

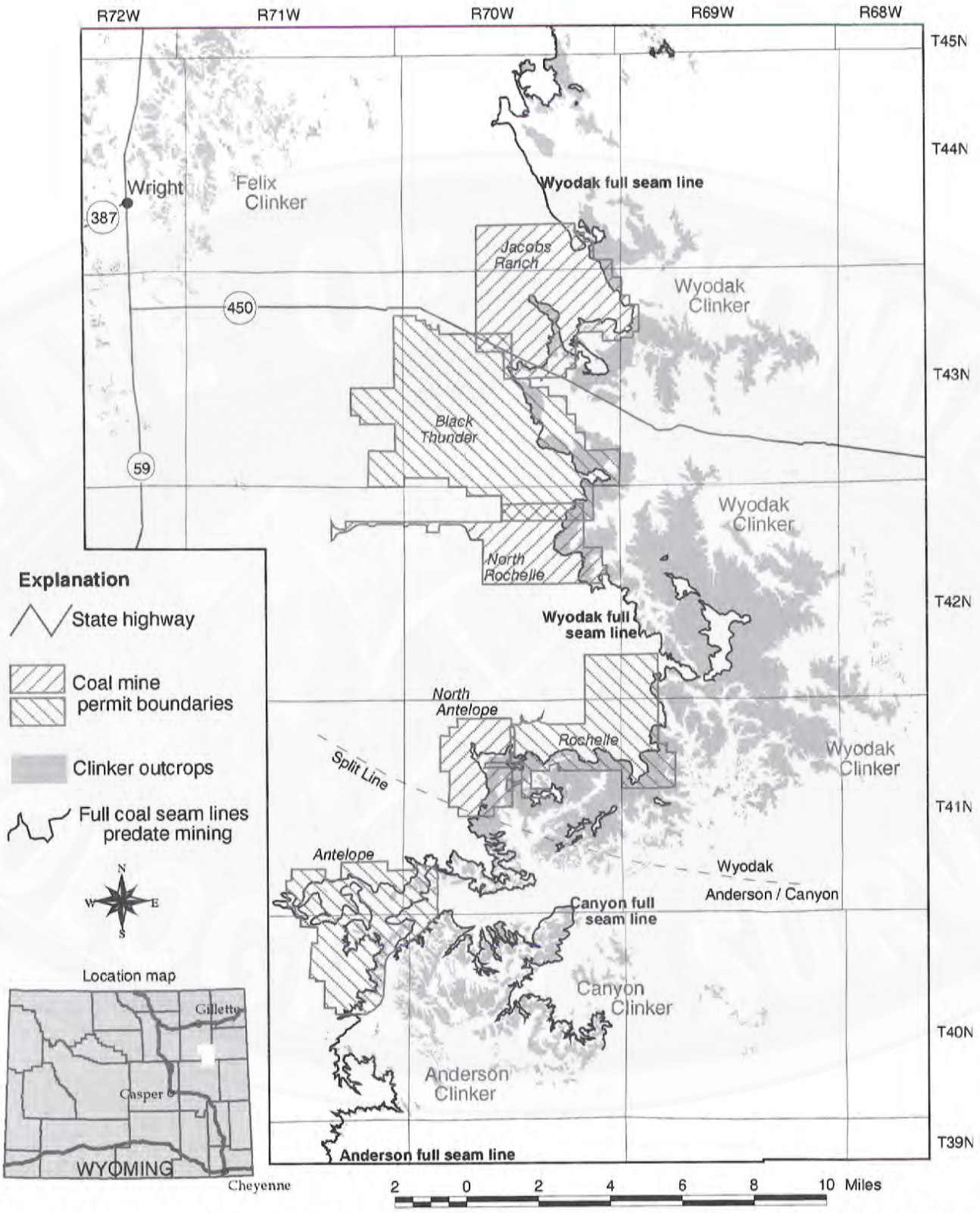


Figure 9. Clinker outcrops in the southern Powder River Basin, Wyoming. Map compiled by Jim Oakleaf of Wyoming Water Resources Center from clinker data provided by the authors.



Figure 10. South-facing view of Lake De Smet, Johnson County, Wyoming. Much of the lake's shoreline is composed of red clinker.



Figure 11. Clinker capping ridges along Boxelder Creek, northeast of Buffalo, Wyoming.

Felix area

Clinker formed by the burning of the Felix and other coal beds in the Wasatch Formation (**Figure 6**) caps a number of buttes and knobs northwest and west of Gillette and south along Highway 59 toward Wright (**Figure 7**). These Wasatch coal beds have formed about 60 mi^2 (160 km^2) of clinker east of the Powder River (Reheis, 1987; Reheis and Coates, 1987).

Powder River breaks

Clinker covers about 40 mi^2 (100 km^2) along the breaks created by downcutting of the Powder River, just south of the Wyoming-Montana border (**Figure 7**). This clinkered area, formed by burning of the Smith, Anderson, Dietz, and Canyon coals in the Tongue River Member of the Fort Union Formation (**Figure 6**), extends north into Montana (Olive, 1957).

Tongue River breaks

West of the Tongue River and northeast of Interstate 90, about 20 mi^2 (50 km^2) of clinker occurs in the Tongue River Member of the Fort Union Formation and the lowermost Wasatch Formation, south of the Wyoming-Montana border (**Figure 7**). This zone extends northward into Montana, where the clinker is even more extensive north and west of Tongue River Reservoir (Heffern and others, 1993). This clinker was produced by burning of the Roland, Smith, Anderson, Dietz, Canyon, and Monarch coals (**Figure 6**), which crop out along the more steeply dipping western flank of the PRB (Barnum, 1975, 1983). The pattern of burning is complicated by a series of en echelon, east-northeast-trending scissor faults that disrupt the eastward dip of the strata.

Age of burning

The clinker present today is highly variable in age. During regional downwasting of the landscape, the local water table has lowered in response to downcutting by streams. A portion of a coal bed that is deprived of ground water is likely to burn quickly, geologically speaking, regardless of ignition source. Thus the clinker nearer the burn line is generally younger than that farther away from the burn line. In addition, thicker coal beds can bake thicker overburden, which is more resistant to erosion, so that the older clinker in place is generally from thick coal beds.

The east side of the PRB illustrates this principle. The Rochelle Hills are capped by clinker of the Wyodak-Anderson coal zone, which is being mined immediately west of the clinker. The beds dip westward 50-100 feet per mile (9-20 m per km), while the landscape drains to the east. As a result, the plateaus and ridges of clinker that extend eastward from the burn line rise higher above the streams and the water table toward the east (**Figure 8**).

C.W. Naeser of the U.S. Geological Survey determined fission-track ages of zircon grains found in clinker from fourteen sites on the ridge north of Little Thunder Creek in the Rochelle Hills east of Wright (Coates and Naeser, 1984). These ages show a trend from about 700,000 years before present (y.b.p.) updip on the east down to less than 18,000 y.b.p. (too young to date using fission tracks) near the burn line downdip on the west.

Although the oldest age determined to date for clinker in Wyoming is 0.7 million years (Ma), this should not suggest that burning started only at that time. Clinker has been forming more or less continuously, geologically speaking, since erosion began to exhume the PRB, probably in Pliocene time. Clinker capping the Little Wolf Mountains in Montana yields fission-track ages as old as 2.8 Ma and clinker capping the divide between the Tongue River and Lame Deer Creek on the Northern Cheyenne Indian Reservation is as old as 1.4 Ma (Heffern and others, 1983). Much of the older clinker has been eroded away and its fragments transported downstream.

Paleomagnetism also can be used to constrain the age of clinker. In both Wyoming and Montana, rocks with fission-track ages in the range of 700,000 y.b.p. or greater show reversed magnetism, documenting Bruhnes and Matuyama age burning and confirming the fission-track ages (Jones and others, 1984).

Effects of clinker on coal mining

Mining of coal near the burn line presents four major challenges for coal mine operators: determining the location of the coal; preventing water inflow problems; maintaining adequate coal quality; and dealing with increased mining difficulty.

Determining the location of the coal

Mines employ three main methods to delineate the burn line: a) surface mapping, augmented by use of color or false-color infrared aerial photos–clinker is red on color photos and gold on false-color infrared; b) closely spaced drilling, on 50- to 100-foot centers with the rows of holes oriented perpendicular to the burn line; and c) magnetometer surveys, run perpendicular to the burn line. **Figure 9** shows the relation of the burn line and clinker to coal mines in the southern PRB. This map was derived using a combination of all three methods to delineate the burn line. The clinker produces irregular magnetic anomalies with short wavelength and high amplitude (Hasbrouck and Hadsell, 1978). Aeromagnetic surveys have revealed that areas of clinker are magnetically very "noisy." The high-frequency component of the magnetics can be used to map clinkered areas, which show as closely-spaced, high amplitude anomalies (Gay and Hawley, 1991). Magnetite, titanomagnetite, and other forms of reduced iron in the clinker chimneys are likely the dominant source of the magnetism (Miller, 1979).

Preventing water inflow problems

Mines on the eastern flank of the PRB require careful planning to avoid water inflow from large saturated clinker bodies updip from the mines (**Figure 12**). The amount of inflow can vary. Some mines excavate to the burn line or even recover partial coal seams underneath clinker without encountering water problems. Others leave a narrow fender of unmined coal between the mine pit and the clinker to prevent flooding from water in clinker updip (Joe Gerlach, Aqua Terra Consultants; Brad Clark, Triton Coal Company; and Roger Mourich, Black Thunder Coal Company, oral communications, 1995). Inflow can saturate the base of the spoil pile downdip and cause stability problems if the water is not pumped out (**Figure 13**).



Figure 12. Frozen waterfall along coal fender at North Rochelle mine. Seepage into mine is caused by ponded water in clinker east of coal fender.

Maintaining adequate coal quality to satisfy contracts

As coal mines approach burn lines, the quality of coal decreases, with losses in volatile matter and carbon content (Sarnecki, 1991). Heating value commonly drops 100 to 200 Btu/lb, while moisture increases 1 to 3 percentage points. Ash content of the mined product increases, partly due to difficulty in removing oxidized "soot" at the top of the coal. Sodium content may decrease 0.5 to 1 percentage point,

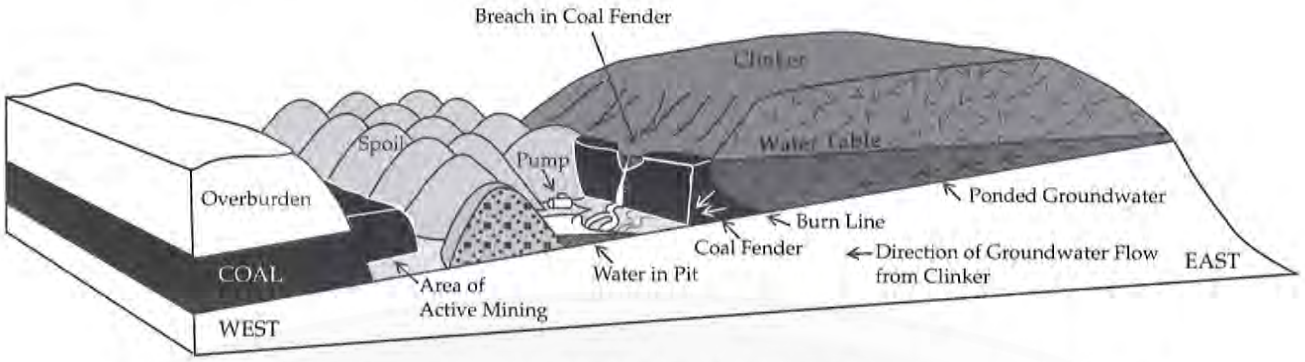


Figure 13. Schematic of groundwater inflow into mine pit from clinker updip, Rochelle Hills, Wyoming.

and sulfur content tends to be more variable near the burn line. These conditions mean that mines must closely monitor the quality of coal near the burn line to ensure that their customers receive a consistent product.

Dealing with increased mining difficulty

Where the top part of a coal bed has burned, several factors complicate removal of the underlying coal. The coal tends to break into finer particles after it is blasted and extracted, which contributes to generation of coal fines. The irregular surface at the top of the coal may require removal of large amounts of soot and coal ash, sometimes on steep slopes. The overlying clinker is harder and more permeable than unbaked overburden, and cannot be drilled and shot because the force of a blast dissipates into the voids in the clinker rather than shattering the rock. This places more wear and tear on shovels. In addition, the stripping ratio (and therefore mining costs) increases because only part of the coal seam is present (Greg Fesko, oral communication, 1995).

Escaping volatile matter and oxidizing coal make coal fires a common occurrence in PRB mines. Piles of finely divided coal, where greater surface area is exposed to moist air, are especially prone to spontaneous combustion. This problem is especially evident on windy days in the spring when standing water is in the pit and the air is humid from rain and melting snow, because water acts as a catalyst for combustion. Sarnecki (1991) suggests that the potential for heat of wetting reactions leading to oxidation and spontaneous combustion can be reduced by maintaining high water levels in the coal near the clinker interface until the coal is mined. Fires from spontaneous combustion in abandoned underground coal mines in the Acme area north of Sheridan are a continuing problem (Dunrud and Osterwald, 1980).

Effects of clinker on oil and gas exploration

Seismic lines for oil and gas exploration should avoid clinker areas. "Near surface clinker beds ... play havoc with seismic interpretation" (Keener, 1985, p. 49). Because of its pervasive fractures, clinker is a very poor medium for transmission of seismic energy. Compressional wave velocities range from 300 to 500 meters per second (m/s), which is lower than velocities in consolidated bedrock (Miller, 1979). When the energy source of a seismic array is placed in or above clinker, it is difficult to obtain a coherent seismic record. Seismograms consist mostly of low-frequency and unpatterned seismic arrivals (Hasbrouck and Hadsell, 1978).

Uses of clinker

In prehistoric times, Native Americans used the porcellanite and glassy forms of clinker for spear points and arrowheads. These materials exhibit conchoidal fracture and were therefore easily flaked and fashioned into stone implements. The material was widely available across the northern Great Plains. Porcellanite tools as old as eleven thousand years, of the Folsom and Goshen styles, have been found (Clark, 1985).

Today, the main use of clinker is to surface unpaved roads in coal mines and rural areas, where better quality gravel is not available. Throughout most of the basin, clinker-red is the typical color of unpaved roads. The clinker provides a well-drained surface but tends to pack poorly and break down easily (Sigsby, 1966). Hundreds, if not thousands, of active and abandoned clinker quarries dot the landscape (**Figure 14**). Loca-



Figure 14. Forest Service clinker quarry at Dull Center, Miller Hills, Converse County, Wyoming.

tions of some of the larger quarries can be found on the industrial minerals and construction materials map of the PRB compiled by Harris and King (1989).

Los Angeles abrasion tests show that the hardness of clinker is variable; the harder and more durable lithologies generally are found closer to the heat source. The dark red to purple or black material found in and near clinker chimneys makes the most durable road metal (Hasbrouck and Hadsell, 1978); however, some of this material is hard on tires. Cement-treated clinker is used as a lightweight aggregate for base courses, but untreated clinker is not widely used as granular base because of lack of durability when wet. Use as granular base is limited to the harder, darker varieties of clinker. Clinker is not used as riprap because it is too lightweight, and generally too highly fractured (Rick Harvey, Wyoming Department of Transportation, oral communication, 1996), but it is used as a topping on embankments to increase the strength of soil. Clinker does not provide good foundations for plant structures or shops because its strength is so variable, and clinker sections may swell and heave differentially during the winter from variable freezing and thawing (Sarnecki, 1991). Clinker was used as aggregate in concrete on many older farms and ranches. In many cases clinker fragments have weathered out leaving pits in the concrete, because it was impossible or impractical to separate out the softer lithologies in preparing the aggregate.

Sigsby (1966, p. 199), in his dissertation on clinker in North Dakota, lists advantages and disadvantages of clinker as an aggregate. Its advantages are availability, abundance, pleasing colors, and ease of development. Because of its low bulk specific gravity, large volumes may be moved in light-duty hauling equipment. Due to its extensive fractures, users of the softer types of clinker may not need explosives or breaking equipment to obtain the material from outcrop. Its disadvantages include small size, large proportion of fines, low compressional strength, low durability, high porosity, and tendency to stain. The high permeability of clinker permits excessive expansion and contraction of moisture during freeze-thaw cycles, leading to rapid breakdown of the rock. The greatest drawback to the widespread use of clinker, however, is the great variation of material properties in outcrop, ranging from moderately soft baked shale and sandstone, to denser porcellanite, to hard paralava.

Sigsby (1966, p. 202) reported that experiments by the North Dakota Highway Department showed that the hardness, strength, and durability of clinker could be dramatically improved for use as a lightweight aggregate by treatment with lime or cement. In recent years this practice has found wide use. The Wyoming Department of Transportation uses clinker mostly for cement-treated base (CTB) for highways. The base is made of a mixture of 92% clinker (1" or 2.5 cm minus size), 6% cement, and 2% fly ash, to which is added 15-30% sandy silt to help the mixture pack down and seal (**Figure 15**). This results in a rigid material with normal strength of 1000 to 1500 psi, although strengths can be as high as 3500 psi. One drawback is that this material shrinks as it dehydrates, causing cracks that extend up through the surface asphalt (Rick Harvey and Ray Burlew, Wyoming Department of Transportation, oral communication, 1996).

Another extensive use of clinker in Wyoming is as chip seal on asphalt roads (**Figure 16**). Clinker between 1/4" and 3/8" (0.6-1.0 cm) diameter is used. In this use, the high porosity of clinker is an asset because it absorbs asphalt and therefore adheres well to the surface. Its light weight reduces the likelihood of broken windshields caused by rock fragments hitting vehicles traveling the highway (Rick Harvey, oral communication, 1996).

Clinker has been used in test sites throughout the PRB for plant mix pavement. However, clinker absorbs up to 6% asphalt while other rock types absorb 1% or less. Normal pavement might contain 5% asphalt, but asphalt mix made with clinker may contain up to 14% asphalt. The high cost of asphalt compared to aggregate makes the cost of clinker pavement prohibitive in most areas. In addition, the light weight of clinker, compared to limestone or granite, means that more clinker must be used to make a given weight of asphalt mix (Rick Harvey and Ray Burlew, oral communication, 1996).

Finely crushed clinker (less than 3/8" or 1.0 cm diameter) has been mixed with salt and used to sand streets in the city of Gillette, as well as on state highways, during the winter. Although clinker is inexpen-

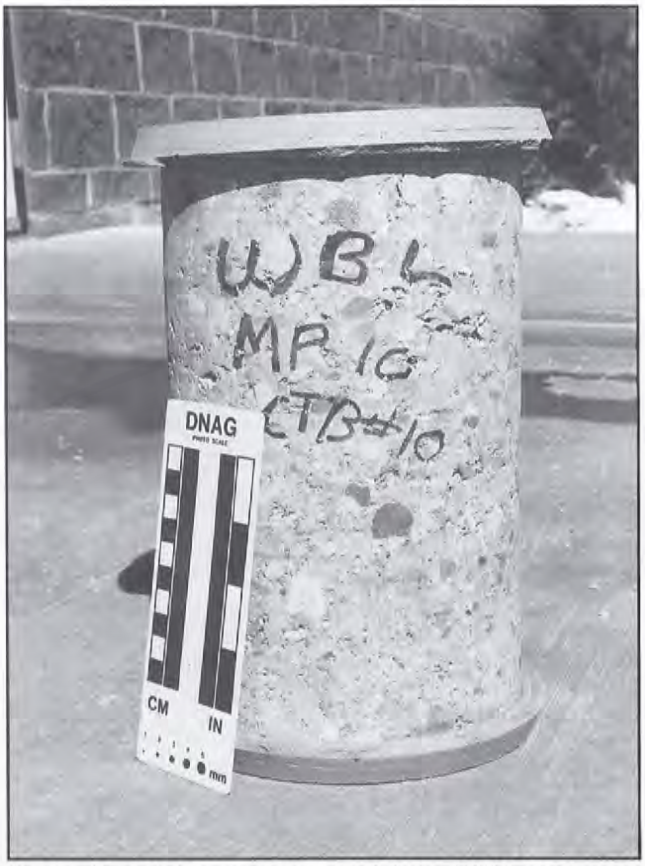


Figure 15. Clinker cement-treated base (CTB) core from Interstate 90, milepost 10 north of Sheridan, Wyoming.



Figure 16. Clinker (darker material) used as experimental chip seal on State Highway 59 near Wright, Wyoming.

sive, it easily breaks down to dust and the contrast between the red color and the gray pavement is an aesthetic nuisance (Gillette News-Record, 1995). The city is considering switching to coal bottom ash from the Wyodak power plant or a sand/limestone mixture (Gillette News-Record, 1996).

Coal mines use clinker for purposes in addition to surfacing mine roads. In New Mexico and Arizona, the McKinley and Black Mesa coal mines use unprocessed clinker in runoff areas, spillways, and ditches to deter erosion and return areas to steady state conditions. One coal company has experimented with clinker as aggregate in both concrete and asphalt. However, the clinker absorbs too much water in the concrete, which detracts from its strength. The company also found that clinker absorbs too much oil in the asphalt (Hoffman, 1996). Clinker is being considered for uses in mine reclamation, such as placement of porous clinker drains within the main recharge zones of reconstructed alluvial and spoil aquifers to help reestablish the aquifer and reduce total dissolved solids in groundwater (Sarnecki, 1991).

One interesting use of clinker is in landslide areas along highways to reduce the weight of the slide. Clinker normally has a specific gravity of about 1.8, significantly lower than the 2.6 figure for most bedrock in the PRB. Reducing the load on the plane of slippage reduces the likelihood of future movement on the slide. In 1994, clinker was used to replace bedrock in a landslide area crossing Interstate 90 between Buffalo and Sheridan (Rick Harvey, oral communication, 1996).

Crushed clinker and clinker boulders are used as decorative aggregate for landscaping throughout eastern Wyoming and Montana and in nearby states (**Figure 17**). Clinker has also been used for lightweight roof covering, texturing in brick manufacture, and sewage filter beds (Sigsby, 1966).

In 1995, seventeen mineral material sales on BLM land in the Wyoming part of the PRB produced 520,164 cubic yards of clinker worth $274,402. Most was purchased by coal companies for surfacing mine roads. An additional free use permit to Campbell County for surfacing county roads produced 112,059 cubic yards (Bob Janssen, BLM, oral communication, 1996). In the Thunder Basin National Grassland, ad-



Figure 17. Clinker boulders used for landscaping at rest stop along State Highway 59 near Bill, Wyoming.

ministered by the U.S. Forest Service (FS), the one 1995 sale produced 291,875 cubic yards of clinker worth $102,155. The amount produced under FS free use permit fluctuates by year depending on the amount of county roadwork: in 1995, there was only one permit for 5 cubic yards, but in 1993, free use permits produced 283,000 cubic yards (David Geer, FS, oral communication, 1996). The BLM and FS allowed coal companies to use additional clinker free of charge if it was necessarily moved in the process of mining coal on a Federal lease. Other clinker was produced from state and private lands.

Conclusions

This paper has discussed the origin and use of the reddish rocks that travelers encounter while driving through the landscape of northeastern Wyoming and southeastern Montana. Many people are unaware that the vast expanses of clinker-capped ridges and buttes would not exist were it not for the natural burning of coal. The huge pits in the PRB where this coal is mined today produce over a quarter of the nation's coal; controlled burning of this coal generates the electricity which powers the homes and workplaces of many Americans. We hope this article helps the reader to better understand the relation between coal and clinker.

Acknowledgments

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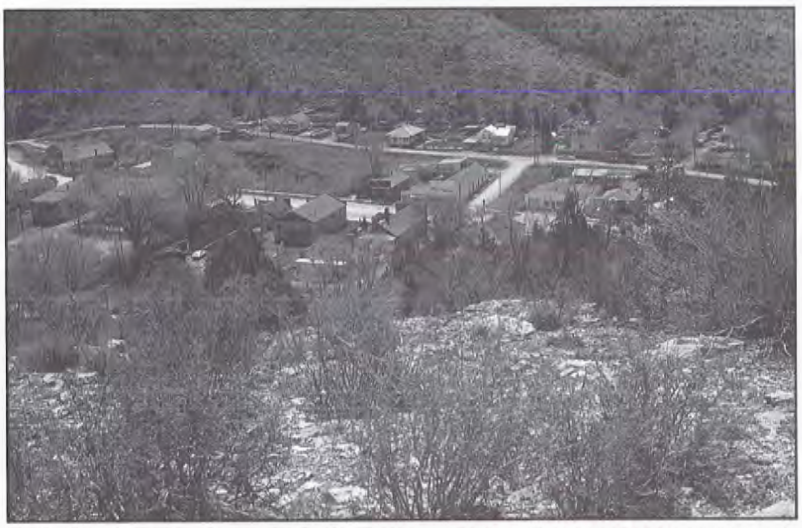
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Hartville townsite, in eastern Wyoming, was the first incorporated town in Wyoming (1858). Originally developed as a copper mining camp, Hartville is better known as the center of iron ore mining in eastern Wyoming for over 100 years, ending in 1981. Shallow copper mineralization in the Guernsey Formation and the uppermost part of the underlying Precambrian schist gives way to hematite resources at depth. The Sunrise and Chicago mines were the largest mines in the Hartville area. (Photograph by R.E. Harris.)

Montana's changing industrial minerals industry

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Abstract

Talc production in Montana has expanded from a modest beginning in the early 1940s to around 400,000 short tons of annual production in recent years, making Montana the leading talc-producing state. Barretts Minerals, Inc. (owned by Minerals Technologies, Inc.) and Luzenac America (owned by RTZ) are the current producers. Bentonite is mined in the extreme southeastern corner of the state by the American Colloid Company and processed in their plant at Colony, Wyoming. In 1982, Continental Lime constructed their Indian Creek plant near Townsend and in 1991, expanded its capacity to 1000 short tons of lime per day with the addition of a second kiln. Basin Electric completed a lime plant at Frannie, Wyoming in 1993 that uses limestone from the Warren quarry just across the border in Montana. In 1995, Cominco American began mining almandine garnet for the abrasive and filter bed markets from dredge tailings along Alder Creek in southwestern Montana. Increased demand for decorative stone for local and out-of-state markets has resulted in numerous independent operations providing stone ranging from Precambrian gneiss to field stone. Interest continues in the development of two vermiculite deposits in southwestern Montana: one near Dillon and the other near Hamilton. Clinoptilolite, from the Grasshopper Creek zeolite deposit near Dillon, is being evaluated for various applications. American Gem has acquired many sapphire deposits in western Montana with plans to produce heat-treated sapphires for mass markets and Gem River is developing a sapphire deposit in southwestern Montana.

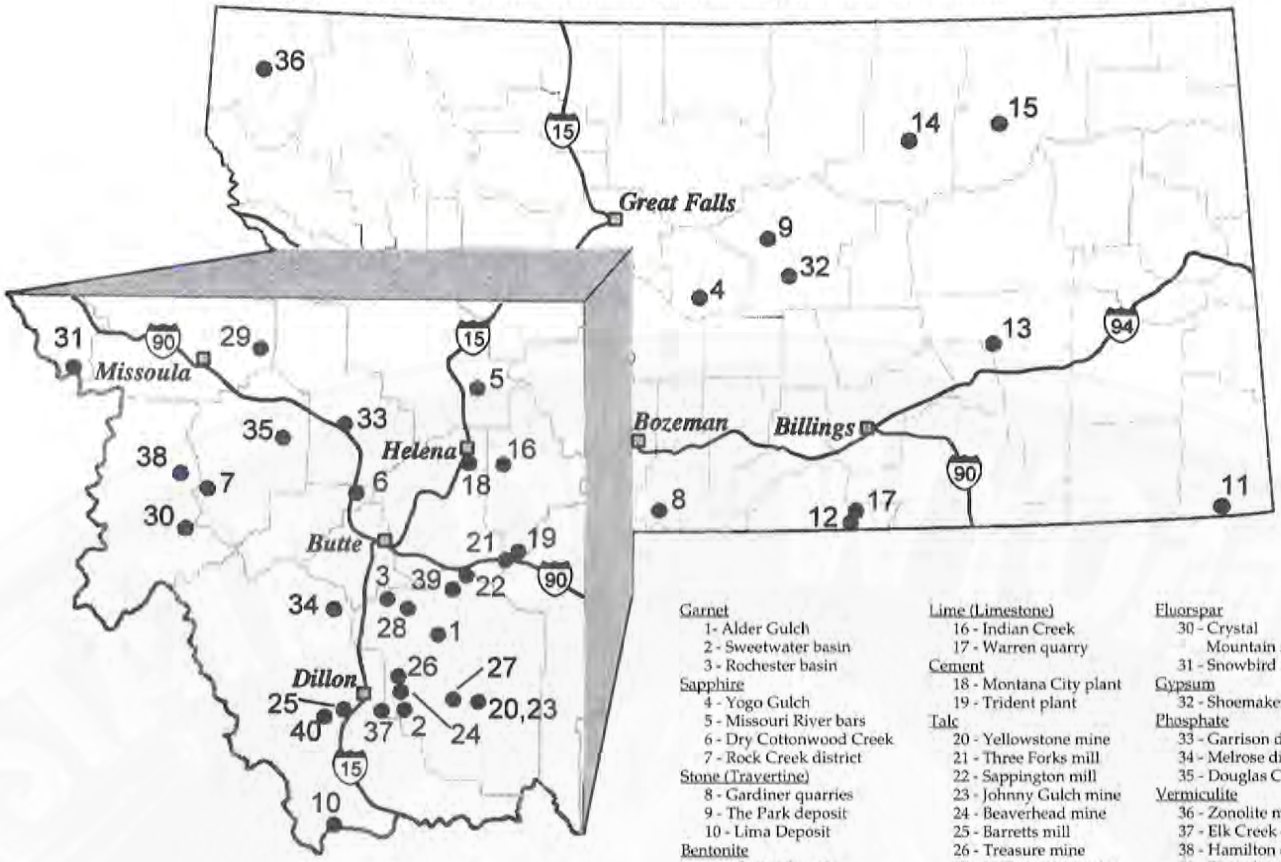
Introduction

As with all industries, the industrial minerals industry in Montana is continually changing. The rate of change of this industry in Montana has increased in the last decade with changes in ownership, new mines and plants, and closures. The purpose of this paper is to document some of these changes for selected commodities before this information is lost. For a more complete list of Montana's deposits of industrial minerals with brief descriptions of the geologic settings of these deposits, see Berg (1990). Because there are generally less than three producers of a specific mineral commodity in Montana, the U.S. Bureau of Mines has not released production figures (to protect individual producers' confidentiality) for most of these commodities mined in Montana.

Mineral commodities with current production

Garnet

The newest industrial mineral operation in Montana is the recovery of almandine garnet from dredge tailings remaining from the placer mining of gold in Alder Gulch, 50 miles (80 km) southeast of Butte (No. 1, **Figure 1**). Between 1899 and 1922, over 2 million ounces of gold were mined from placer deposits along Alder Gulch (Dippold, 1996). The source of the alluvium in Alder Gulch is metamorphic rock of Archean age exposed in the surrounding mountains. Amphibolite, hornblende gneiss, quartzofeldspathic gneiss, and less abundant schist have provided garnet to the alluvial deposits. The dredge tailings in Alder Gulch, which are more than 36 feet (11 m) thick, are reported to contain 4.5% almandine garnet and 0.03 gram/tonne



Garnet
1- Alder Gulch
2 - Sweetwater basin
3 - Rochester basin
Sapphire
4 - Yogo Gulch
5 - Missouri River bars
6 - Dry Cottonwood Creek
7 - Rock Creek district
Stone (Travertine)
8 - Gardiner quarries
9 - The Park deposit
10 - Lima Deposit
Bentonite
11 - Carter County
12 - South of Billings
13 - Vananda
14 - Malta
15 - Brazil Creek
Lime (Limestone)
16 - Indian Creek
17 - Warren quarry
Cement
18 - Montana City plant
19 - Trident plant
Talc
20 - Yellowstone mine
21 - Three Forks mill
22 - Sappington mill
23 - Johnny Gulch mine
24 - Beaverhead mine
25 - Barretts mill
26 - Treasure mine
27 - Willow Creek mine
Chlorite
28 - Antler mine
Barite
29 - Elk Creek (Greenough) mine
Fluorspar
30 - Crystal Mountain mine
31 - Snowbird mine
Gypsum
32 - Shoemaker mine
Phosphate
33 - Garrison district
34 - Melrose district
35 - Douglas Creek mine
Vermiculite
36 - Zonolite mine
37 - Elk Creek deposit
38 - Hamilton deposit
39 - Pony deposit
Zeolite
40 - Grasshopper Creek deposit

Figure 1. Mines, plants, and deposits described in text.

of gold (Austin, 1994). In the spring of 1995, Cominco American, Inc. began construction of a facility to recover garnet and gold. They will sell a product that consists of 90% almandine and in specific size fractions ranging from 6 mesh (3.4 mm) down to 150 mesh (0.1 mm) for the filter bed and abrasive markets, including abrasive jet cutting. The construction phase of this development has been completed and production is now beginning. Because of health concerns, the use of quartz in sandblasting is now prohibited in some states which has caused an increase in the demand for garnet by this industry.

Sweetwater Garnet, Inc. of Absolute Resources has announced plans to begin mining almandine garnet from an alluvial deposit in the Sweetwater basin 14 miles (22 km) southeast of Dillon (No. 2, **Figure 1**). This deposit is reported to contain 8% garnet which was derived from erosion of Precambrian metamorphic rocks exposed in the Ruby Range surrounding this basin. Sweetwater Garnet plans to construct a facility at the deposit for initial processing to be followed by further processing at a plant situated near Dillon. Garnet is also concentrated in Rochester basin, 25 miles (40 km) southeast of Butte (No. 3, **Figure 1**).

Sapphire

Within the last two years the sapphire industry in Montana has changed dramatically with the involvement of two new companies, American Gem, Inc. and Gem River. Montana has a long history of sapphire production beginning in about 1896 with mining of the famous blue stones from the Yogo deposit 45 miles (72 km) southwest of Lewistown in central Montana (No. 4, **Figure 1**). Sapphires were mined from a lamprophyre dike of Tertiary age until 1929 (Zeihen, 1987). Since then several companies have either mined sapphires from the Yogo dike and associated placers, or investigated and promoted the development of these deposits. Sapphire International Corporation began mining sapphires from this dike in about 1970. Roncor Inc. is currently recovering sapphires from some of the previously mined material. In 1984, Vortex Mining discovered a new occurrence of sapphires within a mile of the now inactive American mine in an altered lamprophyre dike similar to the Yogo dike. Vortex is now mining at another site less than 600 feet (200 m) from the American mine. The Yogo deposit is the only developed sapphire deposit in Montana that produces a blue stone of consistent color.

In addition to the Yogo deposit, there are three other major sapphire deposits in Montana, all placer deposits and all without a recognized bedrock source. Stones from these deposits differ in color from those mined from the Yogo deposit; these deposits generally yield stones that are pale shades of blue to green; and stones from each deposit show a variety of shades. Large concentrations of sapphires occur in the so-called "bars" along the Missouri River, east of Helena (No. 5, **Figure 1**). Probably the largest of these deposits of terrace gravel is the Eldorado Bar, which was extensively sampled by American Gem in 1994. French Bar and Spokane Bar, also along the Missouri River, have been mined for sapphires. Fee digging open to the public has been a major activity for these deposits. Farther west, 15 miles (24 km) north of Butte, sapphire deposits along Dry Cottonwood Creek have received recent attention (No. 6, **Figure 1**). In 1995, American Gem, Inc. mined sapphires from their holdings along Dry Cottonwood Creek and Gem River was involved in developing their property, also along Dry Cottonwood Creek.

Alluvial deposits of sapphire occur in several tributaries of Rock Creek in the Sapphire Range, 60 miles (100 km) northwest of Butte, (No. 7, **Figure 1**). For many years buckets of sapphire-bearing gravel from the Gem Mountain deposit were sold to the public. In 1995, American Gem mined this deposit commercially and in 1996 reported that they had mined 1.56 million carats of sapphires from their Gem Mountain and Dry Cottonwood Creek deposits. Sapphires mined from these deposits are heat treated to enhance their color. In a study of more than 75,000 stones from the Rock Creek deposit, it was found that 65 to 70% of the stones were susceptible to heat treatment, changing to attractive blues and yellows from pale shades of blue and green (Emmett and Douthit, 1993). With the successful heat treatment of pale sapphires from the western Montana deposits, new markets have been developed for the commercial mining of these deposits. In addition to these deposits, there are five other alluvial occurrences of sapphires in western Montana (Zeihen, 1987).

Stone

Over 80 years ago, most Montana cities had a local stone quarry and quite possibly a brick plant using local clays. Because of large capital costs and improved quality standards, these industries have largely vanished from the local level. The last brick plant to operate in the state was Lovell Clay Products' plant in Billings which was closed in the late 1970s. During the past 30 years sandstone, limestone, quartzite, dolomitic siltstone, argillite, and a variety of volcanic rocks have been quarried in Montana. However, the only stone producer that has been in continuous operation is Livingston Marble and Granite who, contrary to their name, quarry and finish travertine from their quarries at Gardiner 40 miles (64 km) south of Livingston (No. 8, **Figure 1**). This firm has been in continuous production for 40 years. Livingston Marble and Granite quarries various colors of travertine, ranging from white to ochre, from which they make tiles, ashlar, and custom-fabricated pieces at their Livingston plant. There are two other large deposits of travertine in Montana: one in the central part of the state, about 12 miles (20 km) north of Lewistown, and the other along the Idaho-Montana border, 50 miles (80 km) southwest of Dillon (Nos. 9 and 10, respectively, **Figure 1**). Neither deposit is now being quarried, but travertine has been quarried from the deposit southwest of Dillon for fabrication of facing by an Idaho producer.

In recent years there has been a substantial increase in numbers of independent stone producers. Argillite in the Prichard Formation of the Proterozoic Belt Supergroup is quarried at numerous localities in northwestern Montana. This tan-to-brown argillite can be split into thin slabs to provide attractive stone for patios, floors, and facing. Much of this stone is marketed outside Montana. Water-worn cobbles and boulders, largely of metamorphic rocks found in alluvial deposits in western Montana and known in the market as river rock, are now in demand. Boulders, some large enough to weigh 5 tons, are sold for landscaping. Loose blocks of sandstone of Cretaceous age (field stone) are collected in central Montana and sold for facing. Lichen-covered sandstone (moss rock) also is sold for facing. The current market is largely for stone with natural-appearing or weathered surfaces.

Bentonite

Bentonite has been commercially mined from at least five localities in Montana. The most important locality for continuous production and also for total production is in the extreme southeastern corner of the state in Carter County. American Colloid mines bentonite (No. 11, **Figure 1**) that is processed at their plant near Colony, Wyoming. These deposits are at the western edge of the very important Black Hills bentonite district. Bentonite is mined from beds in the Belle Fourche Shale, of Cretaceous age. Bentonite from these deposits has been used in pelletizing taconite iron ore, bonding foundry molding sand, oil well drilling mud, pet litter, and engineering applications such as pond lining.

Bentonite also is mined by Wyo-Ben, Inc. from beds in the Thermopolis Shale of Cretaceous age close to the Wyoming border 50 miles (80 km) south of Billings (No. 12, **Figure 1**).

Most other bentonite deposits that have been mined in Montana are in the Bearpaw Shale, also of Cretaceous age. This formation is exposed over large areas of eastern and central Montana. In 1966, Hallet Minerals Company began mining bentonite from the Vananda deposit situated 20 miles (32 km) northwest of Forsyth (No. 13, **Figure 1**), in an area well described by the names of the two local creeks, Froze-to-Death and Starved-to-Death Creeks. After partial drying in a rotary kiln, the bentonite was shipped to Minnesota for pelletizing taconite iron ore. Mining bentonite from this deposit was discontinued in 1978.

In the same year (1978) American Colloid began construction of a plant near Malta to process bentonite mined from beds in the Bearpaw Shale, south of Malta (No. 14, **Figure 1**). Bentonite mined from these beds was used in drilling mud and in pelletizing taconite iron ore. Production of bentonite by the Malta plant continued until 1986, when production was discontinued because of diminishing markets and an unfavorable tax situation.

In the late 1960s, an 18-mile (29-km)-long railroad spur was constructed from the Great Northern Railroad main line to the Brazil Creek bentonite deposit, 18 miles (29 km) southwest of Glasgow (No. 15, **Figure 1**). This bentonite deposit, also in the Bearpaw Shale, was initially developed by the Ashland Chemical Company, then acquired by the Archer Daniels Midland Company, and finally by Federal Bentonite, who closed this drying facility in 1980. The market for this bentonite was mainly the taconite pelletizing industry.

Lime

In 1982, Continental Lime began producing lime at their Indian Creek plant northwest of Townsend approximately 30 miles (48 km) southeast of Helena (No. 16, **Figure 1**). Limestone is quarried from the Mission Canyon Formation of the Madison Group of Mississippian age. Addition of a second kiln in 1991 doubled the capacity of this plant to 1000 short tons of lime per day. Lime from this plant is used for flue gas desulphurization, processing metalliferous ores, and in the production of precipitated calcium carbonate (PCC) for the paper industry. A lime hydrator was added in 1994 to hydrate lime for use in asphalt. Continental Lime is owned by Graymont.

In 1993, Basin Electric Cooperative constructed a lime plant at Frannie, Wyoming, just across the Montana border, to supply lime for flue gas desulphurization at the company's coal-fired generating plants. Limestone is quarried at Warren, Montana, 50 miles (80 km) south of Billings (No. 17, **Figure 1**), and trucked 8 miles (13 km) to their Frannie plant, now operated by Dakota Lime, a division of Basin Electric. The quarry is in limestone of the Madison Group and also supplies limestone to Holly Sugar Corporation's plants at Sidney, Montana and Torrington, Wyoming, and Western Sugar Company's Lovell, Wyoming and Scottsbluff, Nebraska plants. The limestone is used to produce lime for the processing of sugar beets.

Cement

Montana has two cement plants supplying both local and out-of-state markets. Ash Grove Cement West, Inc. operates a cement plant at Montana City, 5 miles (8 km) southeast of Helena (No. 18, **Figure 1**). Limestone quarried from the Madison Group a few miles west is used at their plant. This plant was initially constructed by Kaiser Cement in 1962 and was acquired in 1987 by Hanson Ltd. when Hanson acquired Kaiser Cement. A few months later Hanson Ltd. sold the plant to Ash Grove Cement West.

Holnam owns and operates a cement plant at Trident, where the Gallatin, Madison, and Jefferson Rivers join to form the Missouri River, 55 miles (88 km) east of Butte (No. 19, **Figure 1**). This plant was constructed in 1910 by Ideal Cement and acquired by Holnam in 1990. This facility probably has the longest record of production using a local mineral resource in Montana. Some cement from this plant is shipped to Spokane, Washington, where, because of its high uniform quality, it is used in the manufacture of concrete railroad ties. Gypsum, used by both cement producers, is shipped from a source in northern Wyoming, whereas iron ore, also used in small quantities in the manufacture of cement, is mined near White Sulphur Springs, in central Montana.

Talc

In recent years, Montana has been the leading state in the United States in talc production, as measured in volume or in value of talc produced, with annual production around 400,000 tons. Production in 1992 was 390,765 metric tons (U.S. Bureau of Mines, 1993). Montana talc producers have been able to compete effectively because of the high purity of talc in

these deposits and also because most deposits are of sufficient size to be mined efficiently in open pit mines. The lack of tremolite or other amphiboles in Montana talc has also been an important market consideration. Dolomitic marble of Archean age has been locally replaced by talc as a result of hydrothermal activity (Berg, 1991). Markets for talc mined in Montana include cosmetics, paint, plastics, paper, rubber, and ceramics. The paper industry, where talc is used for pitch control, continues to be an important market for Montana talc. Another important market is the ceramic substrate used in catalytic convertors, which is made from a mixture of talc and clay. Essentially, all of substrates used in Japan, North America, and Western Europe incorporate talc mined from Montana deposits.

The largest talc mine in Montana is the Yellowstone mine, situated 20 miles (32 km) south of Ennis (No. 20, **Figure 1**). Sierra Talc began mining talc from what is now known as the Yellowstone mine (**Figures 2** and **3**) in the 1940s with some early production of steatite talc used in the manufacture of insulators. Steatite talc occurs in the upper part of the deposit and could be machined into shapes, which were then fired to make ceramic insulators. Talc from this mine (**Figure 4**) is processed at two mills: the larger at Three Forks built in 1960 (No. 21, **Figure 1**) and a smaller mill at Sappington built in 1985 about 13 miles (21 km) southwest of Three Forks (No. 22, **Figure 1**). In 1964, Sierra Talc was acquired by Cyprus Industrial Minerals, which in turn was acquired by Luzenac America, Inc. in 1992. Cyprus Industrial Minerals installed a mechanical sorting device for the separation of talc from waste dolomite at their Yellowstone mine in 1987. Until the development of this mechanical sorter, essentially all talc produced in Montana was sorted by hand on sorting belts. The mechanical sorter makes use of the difference in coefficient of friction between talc and dolomite. Crushed and sized ore that consists of talc and dolomite is fed onto a revolving disc where talc slides to the outer edge more rapidly than the less slippery dolomite. In December 1991, Cyprus Industrial Minerals obtained a U.S. patent on this mechanical sorter.

Another company, Montana Talc, began construction of a mill at Sappington in 1985 and in 1986 began stripping at their Johnny Gulch mine (No. 23, **Figure 1**) adjacent to the Yellowstone mine. Montana Talc was a joint venture between Nicor Mineral Ventures and Meridian Land and Minerals Company. In 1986, Nicor Mineral Ventures sold their share of ownership to Costain Holdings, Inc. owned by the Costain Group PLC of the United Kingdom. Westmont Mining Inc., a wholly owned subsidiary of Costain, became the sole owner of Montana Talc in 1989. Luzenac America,



Figure 2. A distant view of the earliest mining of talc in the fall of 1941 at what is now the Yellowstone mine (Anderson, 1942). The mine dump is the light-colored area in the lower center of the photograph.

Inc. purchased Montana Talc in May, 1994 and immediately closed Montana Talc's Johnny Gulch open pit mine, citing safety considerations. Montana Talc's Sappington mill, now owned by Luzenac, continues in production. During the last two years Luzenac America, Inc., has spent $7 million on expansion and modernization of the Three Forks mill which now has a capacity of 125,000 tons/year. The Sappington mill has a capacity of 110,000 tons/year.

Luzenac America, Inc. also mines very high-quality talc at the Beaverhead mine in the Ruby Range 16 miles (26 km) east of Dillon (No. 24, **Figure 1**). The Beaverhead mine was operated as an open pit until 1986, when, because of the need for excessive stripping, the decision was made to mine this ore body underground.

Talc mining in the Dillon area began about the same time as that south of Ennis, with Tri State Minerals Company (a division of Southern California Minerals Company) beginning to mine talc in 1943. They constructed a crushing plant at Barretts, 7 miles (11 km) southwest of Dillon, in the same year (No. 25, **Figure 1**). This plant was expanded in 1954 with the addition of equipment for pulverizing talc. In the 1960s, Southern California Minerals Company was sold to Pfizer, Inc., whose Pigments and Metals Division operated the talc mine and mill. Talc is mined from two adjacent ore bodies: the Treasure Chest and the Treasure State, at the Treasure mine on Stone Creek, 16 miles (26 km) east of Dillon in the Ruby Range. In 1990, Pfizer constructed a fully automated beneficiation plant that utilizes wet grinding and flotation at their Barretts mill to allow processing some of their less pure talc ore. This is the first flotation mill for talc to be constructed in Montana. In October 1992, a new company, Minerals Technologies, Inc., was formed



Figure 3. A close up view of the Yellowstone talc mine as it appeared in the fall of 1941 (from Anderson, 1942).

to operate what had been Pfizer's minerals business. In April 1993, Pfizer's remaining partial ownership of Minerals Technologies, Inc. was completely acquired by Minerals Technologies. The Treasure mine and Barretts mill are now owned and operated by Barretts Minerals, a subsidiary of Specialty Minerals, Inc., which is owned by Minerals Technologies, Inc. In 1995, Barretts Minerals expanded their mill with the construction of a warehouse and new laboratory facilities. Barretts Minerals plans to mine talc in the near future from the Regal mine, also an open pit mine in the Ruby Range.

The open pit Willow Creek mine was operated by Resource Processors from 1970 until 1979 when it was closed. Talc mined from this deposit, situated in the Greenhorn Range approximately 20 miles (32 km) southwest of Ennis (No. 27, **Figure 1**), was hand sorted at the mine and then shipped directly to processors. Cyprus Industrial Minerals bought the lease on this property from Resource Processors and subsequently lost this lease in a court decision. The property was later leased to Conagra, and after that, to Southern Talc. In 1986 and 1987 some development work was completed on this property, and a small amount of talc was mined.

Chlorite

The Antler mine, probably the only chlorite mine in the United States, is situated near Silver Star, about 30 miles (48 km) southeast of Butte (No. 28, **Figure 1**). Magnesian chlorite, formed by the replacement of Precambrian quartzofeldspathic gneiss, is used in some of the same applications as talc (Berg, 1983). Mining of this deposit was begun in 1977 by Cyprus Industrial Minerals; the property was subsequently acquired by Luzenac America, Inc., when they acquired Cyprus' talc operations in 1992. Annual chlorite production from this mine is about 15,000 short tons.

Barite

Because the major market for barite mined in Montana is as a weighting agent in drilling fluids, barite production from Montana deposits has been largely controlled by oil and gas drilling activity in the United



Figure 4. 1995 Oblique aerial view of the Yellowstone mine in 1995. Photograph provided by Luzenac America, Inc.

States and Canada. All barite deposits that have been mined in the state occur in slightly metamorphosed rocks of the Proterozoic Belt Supergroup exposed in the western part of the state (Berg, 1988). The Finlen and Sheridan Company began mining barite from the Greenough mine on Elk Creek, 30 miles (48 km) east of Missoula (No. 29, **Figure 1**) in 1951, and in 1952 they completed a mill for processing this barite. Both the mine and mill were sold to the Baroid Sales Division of the National Lead Company in 1956, who closed the mill in 1966 because of a declining market. The mill was subsequently dismantled. With an increase in domestic drilling activity in 1976, the Montana Barite Company constructed a mill in Missoula, where they ground barite mined from the Greenough deposit (also known as the Elk Creek mine), and from two other barite deposits near this mine. They also shipped in some barite from a deposit near Northport, Washington. In 1983, the mill and leases were sold to Nicor Mineral Venture, who sold it to Mountain Minerals several years later. Shortly after this sale the Missoula mill was dismantled. Mountain Minerals continued to mine a small amount of barite for specialty markets from the underground Elk Creek mine for several years, but is now processing only a small amount of barite from the Coloma mine situated 3.5 miles (5.6 km) southwest of the Elk Creek mine for the drilling fluid market.

Mineral commodities with no current production

Fluorspar

Fluorspar was mined from the Crystal Mountain deposit 22 miles (35 km) southeast of Hamilton (No. 30, **Figure 1**) from 1952 until 1973, when mining was discontinued, but processing of previously mined ore at the heavy media mill at Darby continued into the 1980s. Mining of this deposit was begun by the Cummings-Roberts Company and later continued by Darby Spar, Inc. The principal market for fluorspar from this mine was the steel industry. F and S Mining Co. also mined fluorspar from the Snowbird deposit, situated along the Idaho-Montana border 36 miles (58 km) southwest of Missoula (No. 31, **Figure 1**). In 1958, after two years of mining, the deposit was considered mined out (Sahinen, 1962). In 1980, a small amount of acid-grade fluorspar was mined from this deposit in an underground mine.

Gypsum

Undergound mining of gypsum at the Shoemaker mine, 9 miles (14 km) southeast of Lewistown (No. 32, **Figure 1**) was begun in 1916, with the construction of a mill in 1920 (Perry, 1949). In 1928, U.S. Gypsum bought the operation and in 1936 constructed a wallboard plant next to the mine. Gypsum mining for the wallboard plant continued until 1987, when the mine and plant were closed primarily because of the high cost of underground mining and prohibitive transportation costs for shipment of wallboard to markets outside Montana.

Phosphate

The Warm Springs mine, located in the Garrison Phosphate district 56 miles (90 km) southeast of Missoula (No. 33, **Figure 1**), was one of the longest operating mines in Montana and was reported to be the only underground phosphate mine in the Western Hemisphere at the time of its closure in 1993. Cominco American, Inc. began mining phosphate from the Phosphoria Formation of Permian age in 1929 and shipped the phosphate rock to their lead-zinc smelters at Trail and Kimberly, British Columbia, where sulfuric acid, recovered in the smelting process, was used in the manufacture of fertilizer from the phosphate rock. This mine was closed because of the higher cost of mining phosphate underground, as compared to surface phosphate mines. Also, transportation costs made it uneconomical to mine and ship this phosphate rock to Trail and Kimberly, British Columbia. In Garrison, Rocky Mountain Phosphate operated a facility for roasting phosphate rock for use as a feed supplement for cattle, initially using phosphate rock mined in the Garrison Phosphate district. For a short period of time, however, this roasting facility used phosphate feed mined in Florida. This unusual example of almost unbelievable transport of phosphate rock was made possible by low freight rates resulting from back hauls. This plant was closed in 1976 and has since been dismantled.

Phosphate was mined underground in the Melrose Phosphate district for production of elemental phosphorus at Stauffer Chemical Company's plant at Silver Bow 23 miles (37 km) southwest of Butte (No. 34, **Figure 1**). In 1968, these underground mines were closed in favor of shipping phosphate rock by unit train from the Wooley Valley mine, near Soda Springs, Idaho, to the Silver Bow plant. In 1987, Rhone Poulenc purchased the phosphate plant at Silver Bow and in 1995, the plant was put on standby because the decision was made to provide customers with elemental phosphorus made on contract at FMC's Pocatello, Idaho plant. FMC's larger plant can produce phosphorus more economically than the Silver Bow plant.

Phosphate was mined at the Douglas Creek mine, 22 miles (35 km) northwest of Deer Lodge (No. 35, **Figure 1**) from 1963 until 1968, when the mine was closed. Montana Phosphate Products Company constructed a flotation plant near the mine for beneficiation of the ore prior to shipment to Kimberly and Trail, British Columbia for fertilizer manufacture.

Vermiculite

A major source of vermiculite in the United States for many years was the Zonolite mine, 8 miles (13 km) northeast of Libby (No. 36, **Figure 1**). Mining of this large vermiculite deposit began in 1925 (Perry, 1948) and continued until 1990. Some stockpiled ore was processed after closure of the mine, but the mine has since been reclaimed and the mill dismantled. Declining demand by the insulation markets was cited as the major reason for closing this mine, which had not been producing at capacity for several years. A contributing factor cited by some in the decision to close the mine is the occurrence of small concentrations of tremolite in the ore. Although removed from the vermiculite in processing, the presence of tremolite presented a constant concern.

A small amount of vermiculite was mined from the Elk Creek deposit 13 miles (21 km) southeast of Dillon (No. 37, **Figure 1**). In 1990 a mill was constructed at this deposit, and vermiculite was mined and processed during 1990 and 1991. This deposit has been extensively drilled and is being further evaluated. The Hamilton deposit, 13 miles (21 km) east of Hamilton (No. 38, **Figure 1**) has been permitted after much discussion and controversy. Another vermiculite deposit situated on the northeastern flank of the Tobacco Root Mountains, 35 miles (56 km) southeast of Butte (No. 39, **Figure 1**) near the small town of Pony and usually referred to as the Pony deposit, has not been thoroughly evaluated.

Zeolite

A deposit of clinoptilolite is situated along Grasshopper Creek 12 miles (20 km) southwest of Dillon (No. 40, **Figure 1**). The Montana Bureau of Mines and Geology has been conducting research on the effectiveness of clinoptilolite from this deposit and from other deposits in removing metals from water. In 1994 Montana Clino-Z mined a 100-ton sample from the Grasshopper Creek deposit for evaluation.

Conclusions

As is generally well recognized, transportation costs to large market areas, notably the West Coast and Midwest markets will continue to be a deterrent to the development of many deposits of industrial minerals. It appears that those minerals with relatively high values, such as some of the specialty minerals, will have the most success in competing in distant markets. The minerals with low market values can generally only compete effectively for local markets, where freight cost is not a serious hindrance.

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STATE OF WYOMING
1933
GEOLOGICAL SURVEY

Industrial mineral resources of Nebraska

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Abstract

In Nebraska "industrial minerals" is used as a group term that includes both nonmetallic and metallic minerals but excludes oil and gas. Of the 705 active quarries, pits, and mines operating at the present time, 8 are clay and shale pits, 25 are sandstone quarries, 21 are limestone quarries, 650 are sand and gravel pits, and one is a uranium mine. Industrial mineral resources and production are distributed throughout the state and are marketed as agricultural lime, cement (masonry and portland), clay brick and tile, construction sand and gravel, gemstones, lime, limestone, expanded perlite, talc (ground), uranium (yellowcake), and exfoliated vermiculite. Several other industrial minerals have been produced in the past and others may be produced in the future. Some of the source rocks and minerals include: barite, bentonite, chalk, diatomaceous earth, flint, gypsum, ocher, potash, quartzite, salt, volcanic ash, and zeolite. The value of Nebraska's industrial mineral products in 1995 was approximately $150 million. Although their value has increased substantially over the last 10 years, quantities produced have remained more or less constant. Most of the industrial mineral products are used within Nebraska, but some are shipped out of the state.

Introduction

"Industrial minerals" is used in Nebraska as a group term that includes nonmetallic and metallic minerals but excludes oil and gas (Burchett, 1991, revised 1995). Today's active production of nonmetallics in the state consists of cement, chalky limestone, clay shale, gemstones, lime, limestone, sand and gravel, sandstone, and siltstone. The only metallic mineral mined is uranium (Burchett and Eversoll, 1994a). During 1995 there were eight clay or shale pits; three lime plants; 21 limestone quarries; approximately 650 sand, gravel, silt or siltstone pits; 25 sandstone pits, and one active uranium mine (Burchett and Eversoll, 1994b; **Figure 1**). Generalized information and regulations pertaining to industrial minerals in Nebraska are summarized by Burchett (1993a).

Geologic setting

The industrial mineral resources (Burchett, 1987, 1993b) of Nebraska are associated in various ways within geologic (Burchett, 1986; Burchett and Pabian, 1991, revised 1992; **Figures 2** and **3**) and tectonic provinces (**Figure 4**). Nebraska's surficial geology is composed mostly of loess (wind-blown silt), dune sand, and alluvium. Glacial deposits are limited to eastern Nebraska. All of the unconsolidated deposits are of Quaternary age and are less than 1.7 million years old.

Underlying the unconsolidated deposits is partly to wholly consolidated bedrock consisting of sedimentary deposits of Cambrian, Ordovician, Silurian, Devonian, Mississippian, Pennsylvanian, Permian, Triassic, Jurassic, Cretaceous, and Tertiary ages. Of these, only rocks of Quaternary, Tertiary, Cretaceous, Permian, and Pennsylvanian age crop out within the state (**Figures 2** and **3**).

Nebraska can be subdivided into several major geologic and tectonic provinces (**Figure 4**). Arches

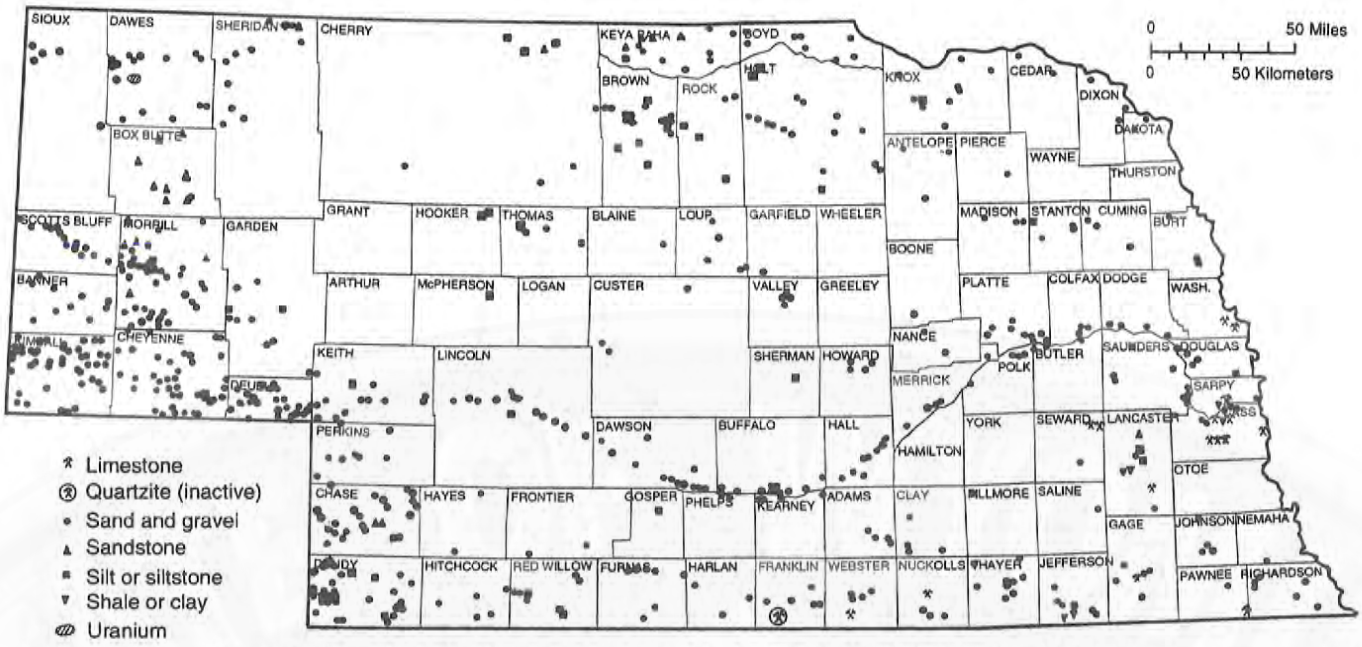


Figure 1. Locations of active quarries, pits, and mines for industrial minerals in Nebraska.

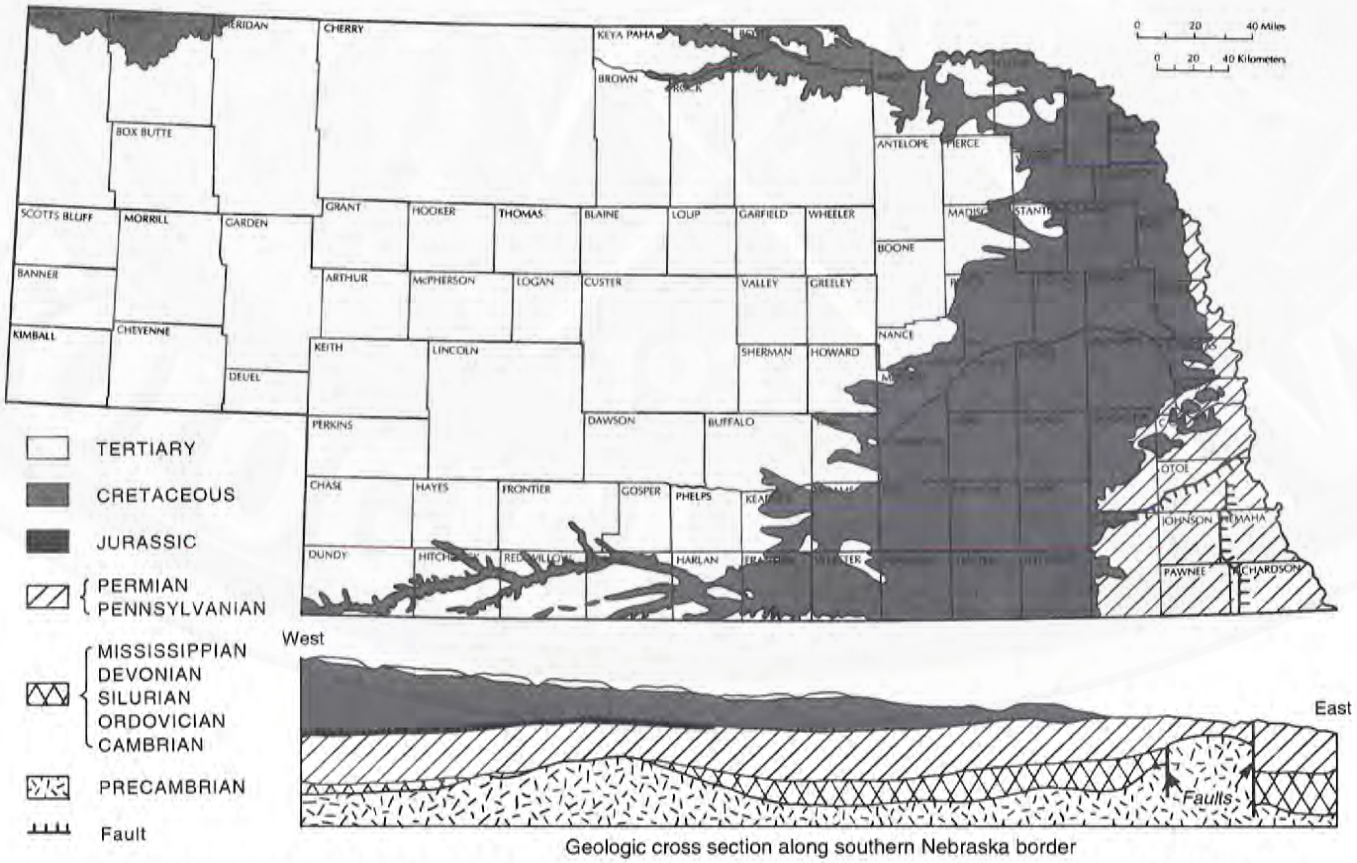


Figure 2. Geologic bedrock map and geologic cross section of Nebraska (from Burchett, 1986; Burchett and Pabian, 1991, revised 1994).

AGE	GEOLOGIC TIME UNITS		ROCK TYPES	MINERAL RESOURCES AND PRODUCTS
MILLIONS OF YEARS AGO	CENOZOIC (recent life)	QUATERNARY (Recent and Pleistocene)	Glacial till, silt, clay, sand, gravel, volcanic ash.	Agricultural soil, water, sand and gravel, volcanic ash.
1.6		TERTIARY	Sandstone, siltstone, clay, gravel, marl, volcanic ash.	Agricultural soil, water, sand and gravel, volcanic ash, riprap and uranium.
66	MESOZOIC (middle life)	CRETACEOUS	Chalk, chalky shale, dark shale, varicolored clay, sandstone, conglomerate.	Water, oil and gas, cement, brick, agricultural lime, and other construction materials.
138		JURASSIC	Subsurface only. Sandstones and shales.	
205		TRIASSIC		
240	PALEOZOIC (ancient life)	PERMIAN	Shale, limestone, dolomite, gypsum, anhydrite sandstone, siltstone, chert.	Water agricultural lime, oil road rock, riprap.
290		PENNSYLVANIAN	Limestone, shale, sandstone, coal.	Oil, cement, brick, concrete aggregate, lightweight aggregate, road rock, agriculture lime, riprap, water.
330		MISSISSIPPIAN	Subsurface only. Limestone, dolomite.	Oil, water.
360		DEVONIAN	Subsurface only. Dolomite, gray shale.	
410		SILURIAN	Subsurface only. Dolomite.	
435		ORDOVICIAN	Subsurface only. Dolomite, sandstone, shale.	
500		CAMBRIAN	Subsurface only. Dolomite, sandstone.	
570	CRYPTOZOIC (hidden life)	PRECAMBRIAN	Subsurface only. Granite, other igneous rocks, and metamorphic rocks.	
5000?				

Figure 3. Geologic time chart of Nebraska (from Burchett, 1986; Burchett and Pabian, 1991, revised 1994).

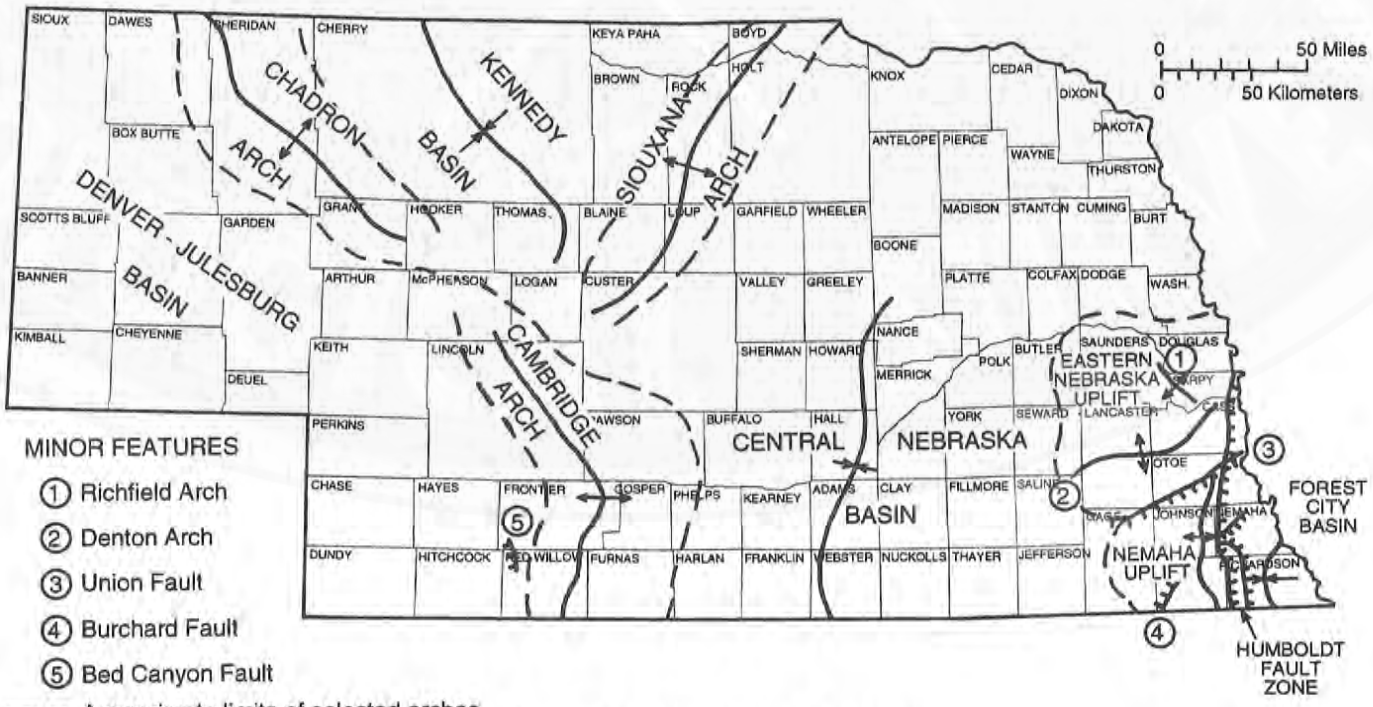


Figure 4. Principal structural features in Nebraska.

and basins associated with these provinces are important because they affect where minerals occur. The western (Panhandle) and southwestern parts of the state lie within the Denver-Julesburg basin. The Chadron arch which trends south-southeastward across the western part of the state, joins the Cambridge arch in west-central Nebraska. In north-central Nebraska the Kennedy basin is limited in extent on the west by the Chadron arch and on the east by the northeast-southwest trending Siouxana arch. The Central Nebraska basin comprises much of central and some of eastern Nebraska. This basin is separated from the Forest City basin in extreme southeastern Nebraska by two major uplifts–the Eastern Nebraska uplift and the Nemaha uplift.

Cement

Cement is a product that when mixed with sand and gravel or limestone aggregate plus water becomes hard like stone and is called concrete. Portland cement has properties allowing it to harden under water. It consists of a controlled mixture of lime, silica, and alumina that forms clinkers when burned in a rotary kiln. Then those clinkers are ground and mixed with gypsum to produce the cement. Portland cement was patented by Joseph Aspdin of Leeds, England, in 1824 and the name "Portland" was chosen because the hardened cement resembled a building stone quarried on the Isle of Portland off the southern coast of England.

Established in 1873 near Beatrice, Nebraska, the first commercial cement plant in the state used limestones and shales of Permian age (Burchett, 1991a). It was forced to close several years later. Several other cement plants operated between 1873 and 1913 but were small and short lived. In 1913, a cement plant at Superior, Nebraska, used Niobrara Chalk and Carlile Shale of Cretaceous age (**Figure 5**). This plant became inactive in 1986. Another plant, established in 1928 at Louisville, Nebraska, uses limestone and shale of Pennsylvanian age and clay of Cretaceous age. Operated by Ash Grove Cement Company and modernized several times, this plant still is operating.

Construction is the most important and principal use of cement, which is used to make pavement, to join bricks, and to pour into forms to build walls. Production of both masonry and Portland cement has increased so much over the last few years that the value of cement accounted for about one-half the total value of non-fuel mineral production in Nebraska.

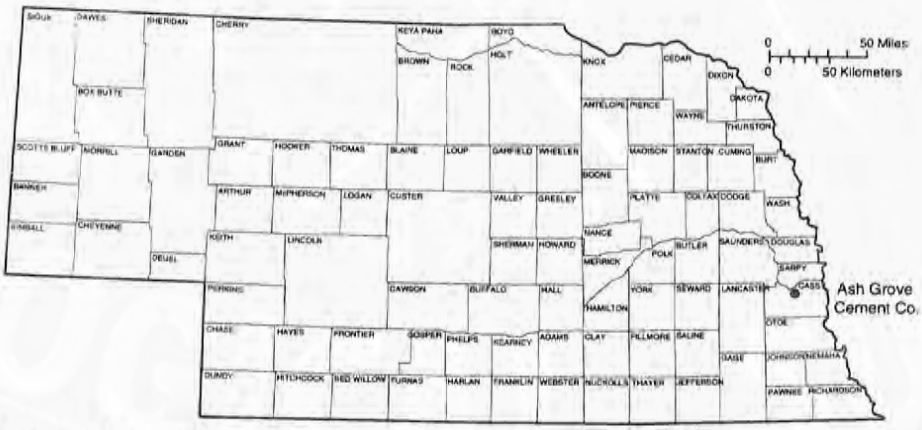


Figure 5. Locations of cement plants in Nebraska. Open circles indicate inactive or abandoned plants, solid dots indicate active plants.

Clay and shale

Clay is the earthy, fine-grained product of the mechanical and chemical break-down of parent rocks and is composed mainly of the minerals illite, montmorillonite, kaolinite, and chlorite. Shale is a sedimentary rock that results from compaction of clay or silt and generally is thin-bedded or laminated.

Six general classifications of clay and shale are recognized: ball clay, bentonite, fire clay, fuller's earth, kaolin, and common clay or shale. Of these, Nebraska has deposits of bentonite, fuller's earth, and common clay or shale (Burchett, 1990c; **Figure 6**). Although present at many places across Nebraska, only the

common clay or shale is produced commercially in the state (**Figure 7**).

In Nebraska, clays and shales are used principally to make brick, tile, and cement. Other uses may become possible.

Clay and shale production in Nebraska during 1995 was estimated at 243,000 tons (**Figure 8**). Nebraska's largest clay and shale producing year was 1985 when 244,000 tons were produced. As of January 1, 1996, a cumulative total of 8,133,315 tons of clay or shale had been produced in Nebraska.

The estimated value of clay and shale production in Nebraska during 1995 was $1,025,000 (**Figure 9**). Production in Nebraska was greatest in 1990 and was valued at $1,685,000. As of January 1, 1996, the cumulative value of clay and shale production in the state was $51,425,669.

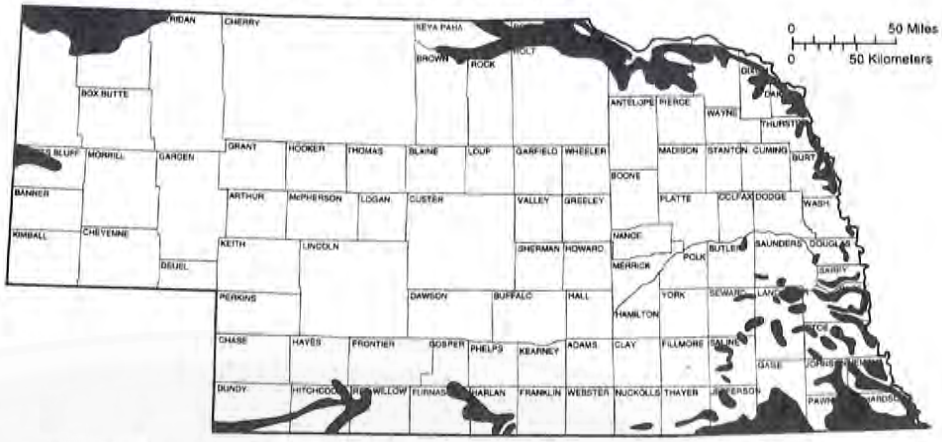


Figure 6. Locations of clay or shale deposits at or near the surface in Nebraska.

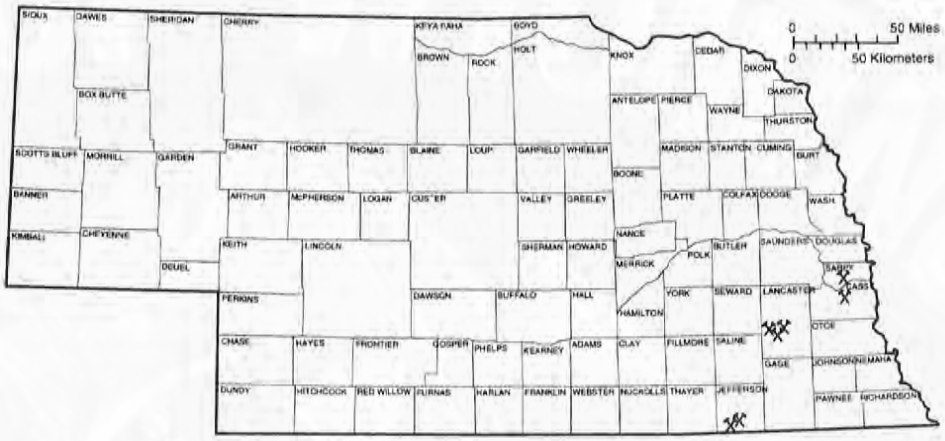


Figure 7. Locations of active clay or shale pits in Nebraska.

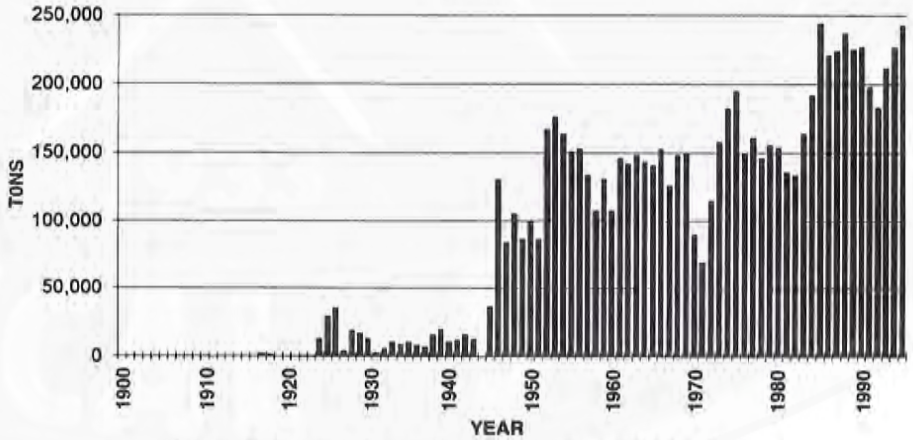


Figure 8. Annual clay or shale production in Nebraska.

Gemstones

Gemstones is a "catch-all" or group name in Nebraska for minerals, rocks, and gemstones. A mineral is an inorganic substance having characteristic physical properties, a definite chemical composition, is homogeneous, and occurs naturally. Gemstones also occur naturally and mostly are colorful, durable, and take a high polish. The value of gemstones in Nebraska in the last 5 years is estimated at $1000 per year (**Figure 10**).

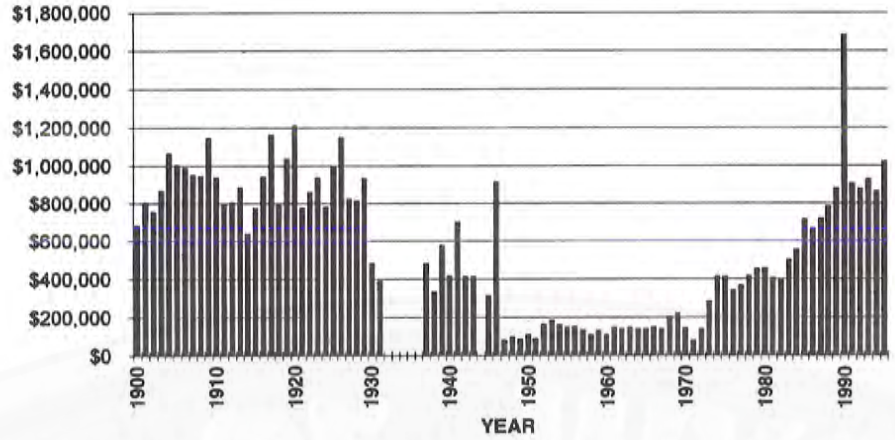


Figure 9. Annual values of clay or shale production in Nebraska.

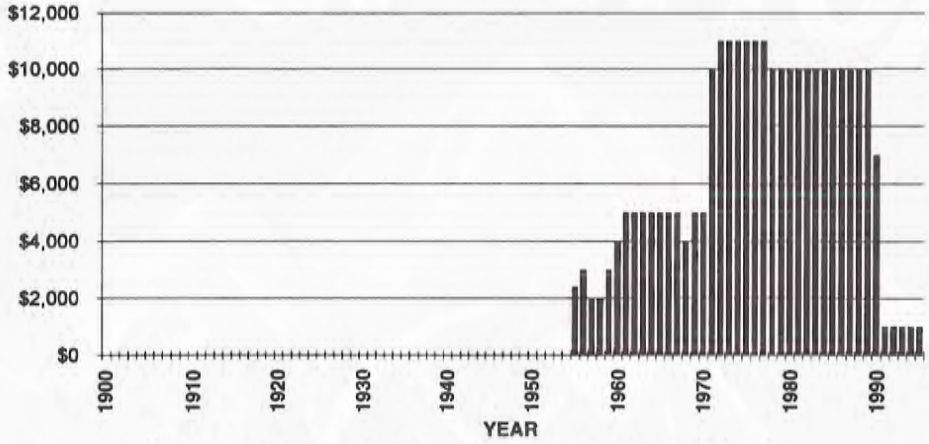


Figure 10. Annual values of gemstone production in Nebraska.

Lime

Western Sugar Company in western Nebraska produces lime from limestone imported from their quarry in Wyoming. The lime is used mainly in sugar refining.

Production of lime in 1995 is estimated at 14,000 tons (**Figure 11**) and had a value of $668,000 (**Figure 12**).

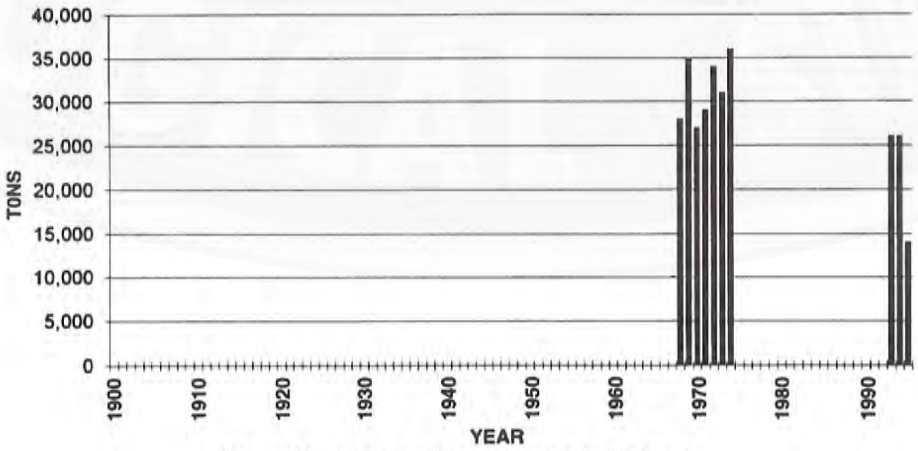


Figure 11. Annual lime production in Nebraska.

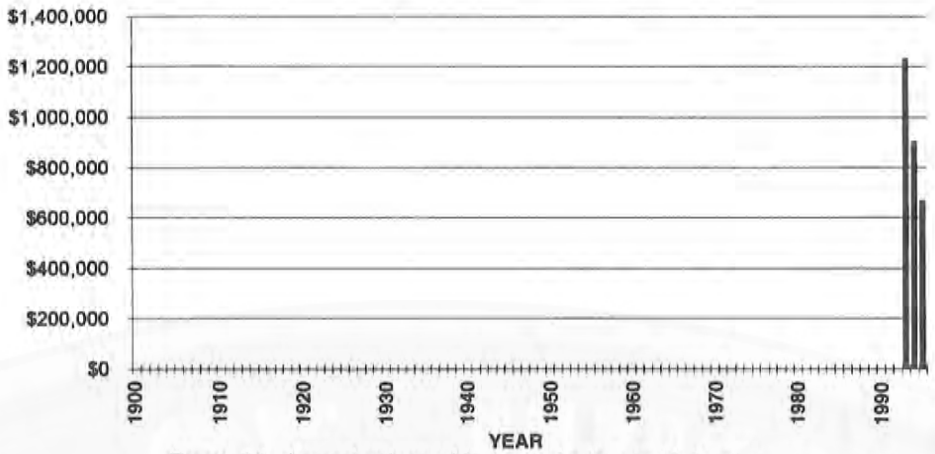


Figure 12. Annual values of lime production in Nebraska.

Limestone

Limestone is a sedimentary rock composed primarily of the mineral calcite (calcium carbonate, $CaCO_3$). Formed by compaction of calcareous ooze that accumulated on the bottom of ancient seas, limestone commonly includes whole or fragmented shells of marine animals that died and their hard parts not wholly dissolved before being incorporated in the ooze.

Limestone crops out mainly in the eastern and southern parts of the state (Burchett, 1990g; **Figure 13**). In the eastern area more than 30 different layers of limestone of Pennsylvanian and Permian age are interlayered with shale or siltstone. Limestone of Cretaceous age is quarried near Garland in Seward County and near Nelson in Nuckolls County for agricultural lime. Cretaceous age limestones were used in the past to manufacture cement near Superior in Nuckolls County. Tertiary age deposits of poor quality limestone "mortar beds" are quarried and used for unpaved road surfacing materials in the northern and western parts of the state.

Although limestone is present beneath most of Nebraska, it is too deeply buried to be quarried or mined profitably at the present time except in eastern Nebraska. The main quarries in Nebraska are located near Weeping Water in Cass County, near Louisville in Cass and Sarpy Counties, and near Ft. Calhoun in Washington County (**Figure 14**).

Limestone has many uses in Nebraska. Crushed limestone is used in the manufacture of cement, as aggregate in concrete and asphalt, for road surfacing, as railroad ballast, and in the manufacture of chemicals, lime, fillers, extenders, feed supplements, and agricultural lime. Sized limestone blocks are used for riprap, retaining walls, and walls of buildings. Currently 13 crushing plants are producing an estimated 7,275,000 tons of limestone per year (**Figure 15**). As of January 1, 1996 a cumulative total of 195,173,954 tons of limestone has been produced in the state.

The estimated economic value of the crushed limestone in 1995 was $39,600,000 (**Figure 16**). As of January 1, 1996, the cumulative value of limestone produced in Nebraska was $618,324,496.

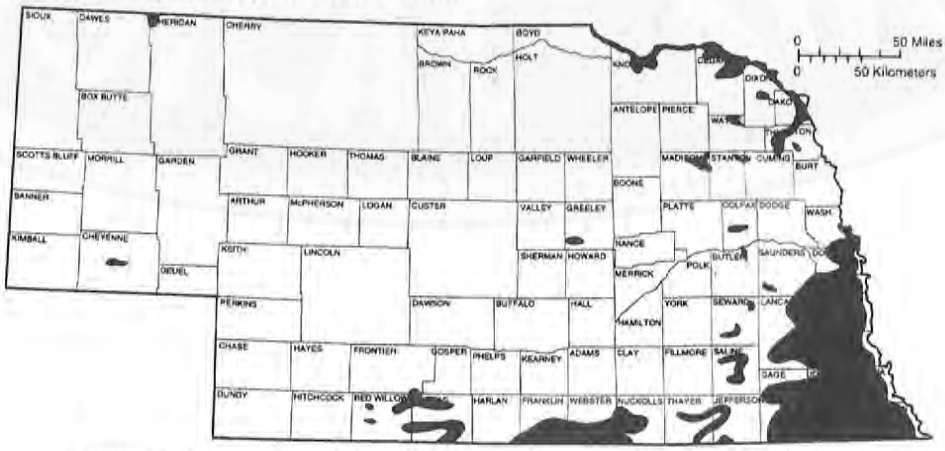


Figure 13. Locations of limestone deposits at or near the surface in Nebraska.

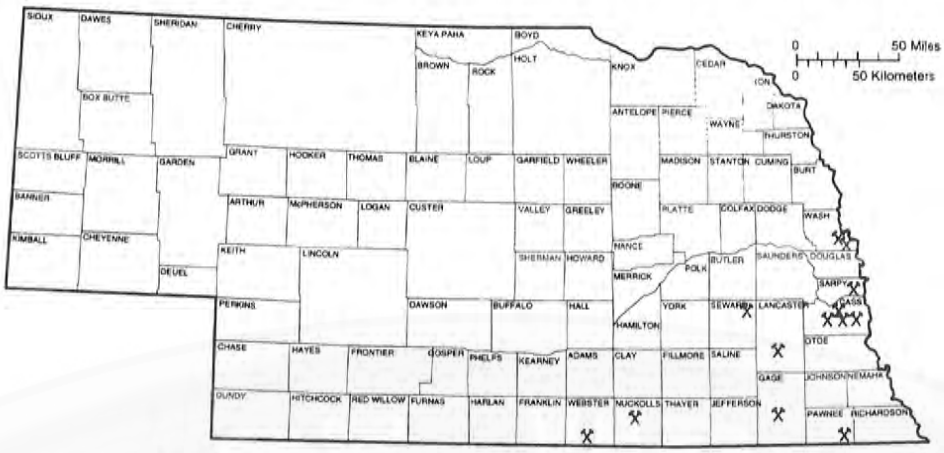


Figure 14. Locations of active limestone quarries in Nebraska.

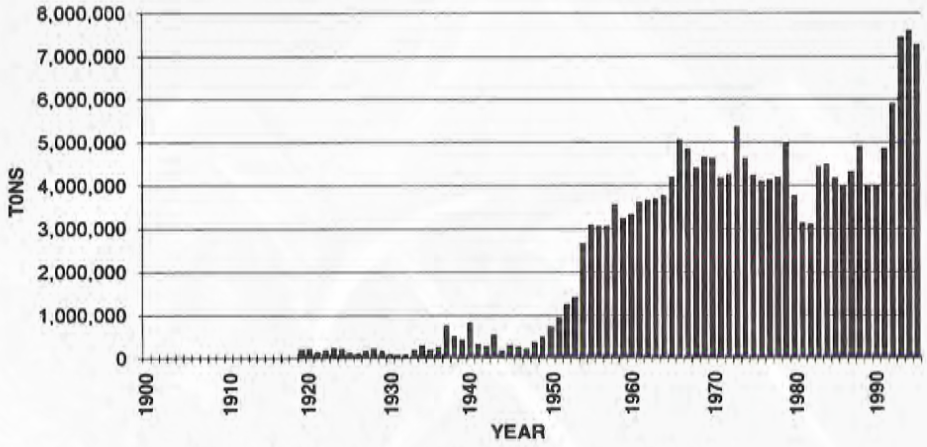


Figure 15. Annual limestone production in Nebraska.

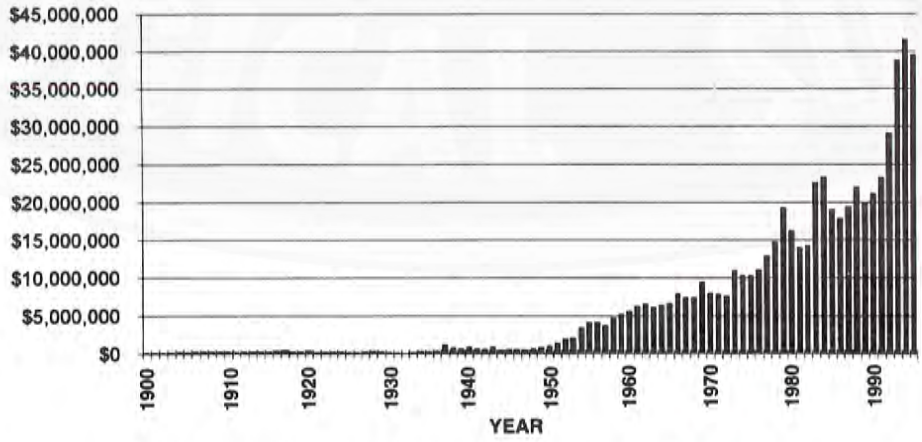


Figure 16. Annual values of limestone production in Nebraska.

Sand and gravel

Sand and gravel deposits consist primarily of quartz and other mineral grains derived from the disintegrating or weathering of rocks. The terms sand and gravel denote particle size. Sand is 1/12 of an inch in diameter (about the size of the head of a match) or less. Gravel is larger than sand and ranges in size from a head of a match to a tennis ball. Cobbles and boulders are even larger.

Deposits of sand and gravel in Nebraska are mostly unconsolidated, have been transported and deposited by water, wind, or ice, and mainly are beneath valley floors. However, some deposits occur in buried valleys beneath upland areas (Burchett, 1990k; **Figure 17**).

Uses of sand and gravel are mainly for construction purposes. Minor amounts of sand are sold for traction and sand blasting.

Estimated sand and gravel production from approximately 650 pits in Nebraska (**Figure 18**) was 17,637,000 tons during 1995 (**Figure 19**). That was the largest sand-and-gravel-producing year, and the estimated value was $55,200,000 (**Figure 20**). As of January 1, 1996, a cumulative total of 641,179,905 tons of sand and gravel has been produced in Nebraska.

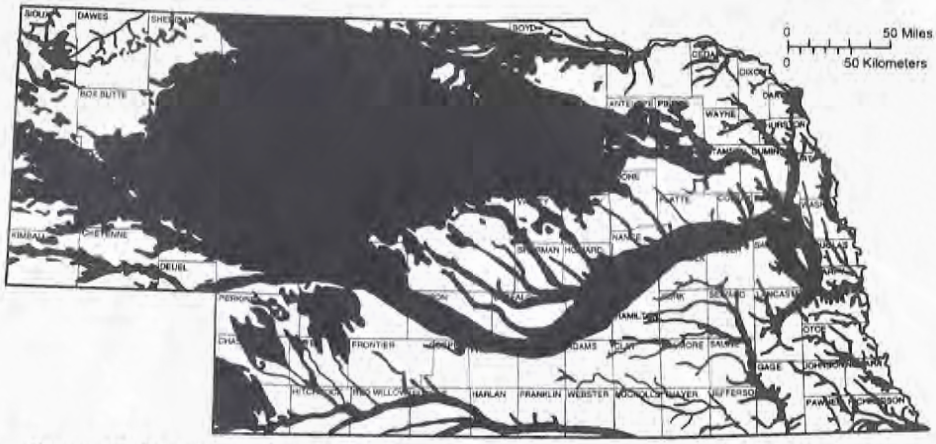


Figure 17. Locations of sand and gravel deposits at or near the surface in Nebraska.

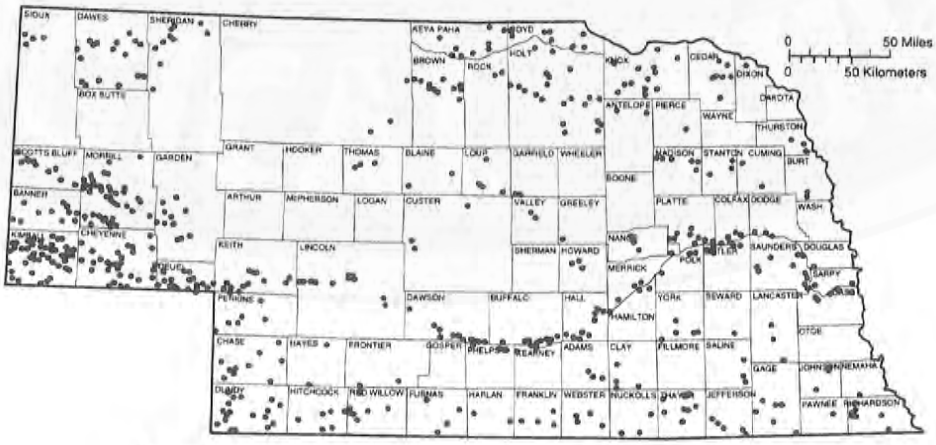


Figure 18. Locations of active sand and gravel pits in Nebraska.

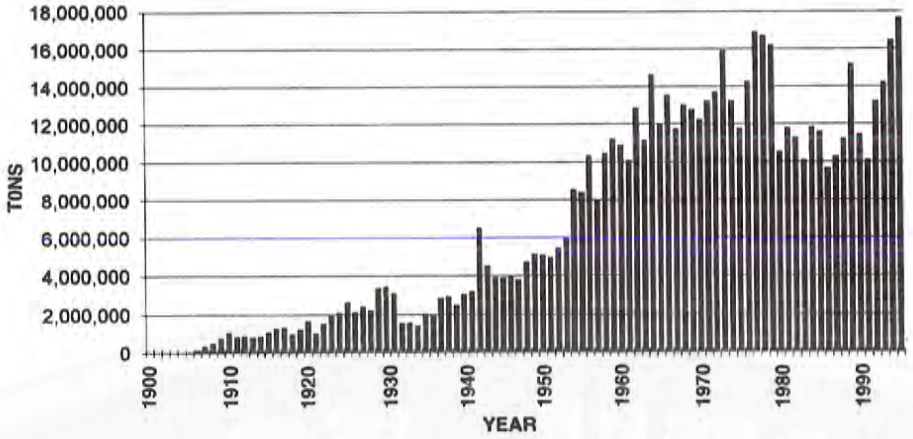


Figure 19. Annual sand and gravel production in Nebraska.

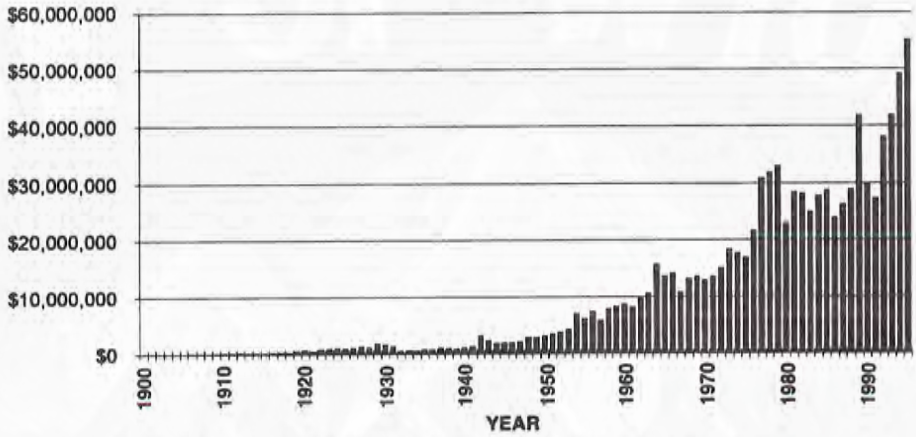


Figure 20. Annual value of sand and gravel production in Nebraska.

Sandstone

Sandstone is rock consisting mainly of quartz and other mineral grains derived from the weathering of disintegrating rocks and then cemented by silica, lime, or iron oxide.

Most sandstones in Nebraska are soft, impure, and interbedded with differing amounts of silt, clay, shale, or limestone (Burchett, 1990l; **Figure 21**). Those sandstone beds close enough to the land surface to be quarried economically range in geologic age from Pennsylvanian to Tertiary. Most of the quarries in eastern Nebraska are in the Dakota Sandstone of Cretaceous age, and those in the north-central and

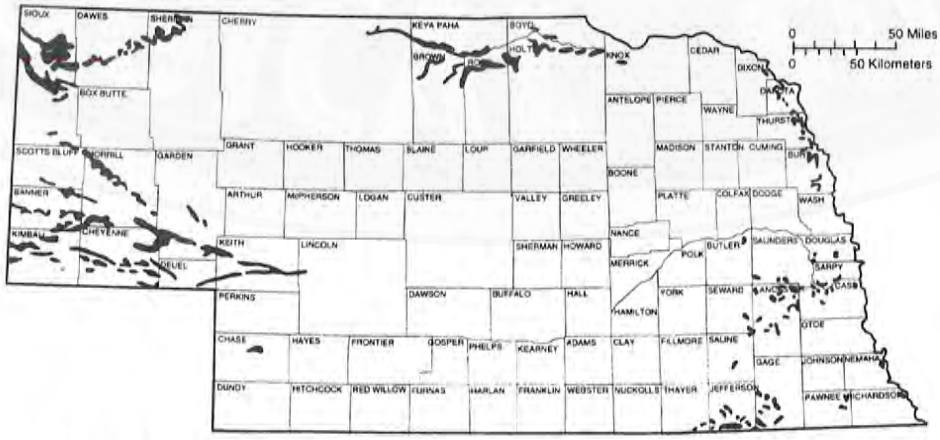


Figure 21. Locations of sandstone deposits at or near the surface in Nebraska.

western parts of the state are in sandstones of Tertiary age (**Figure 22**). During 1995 sandstone was produced in seven counties in Nebraska.

Uses of sandstone in Nebraska are mainly for construction purposes such as fill under concrete or crushed for surfacing of unpaved roads.

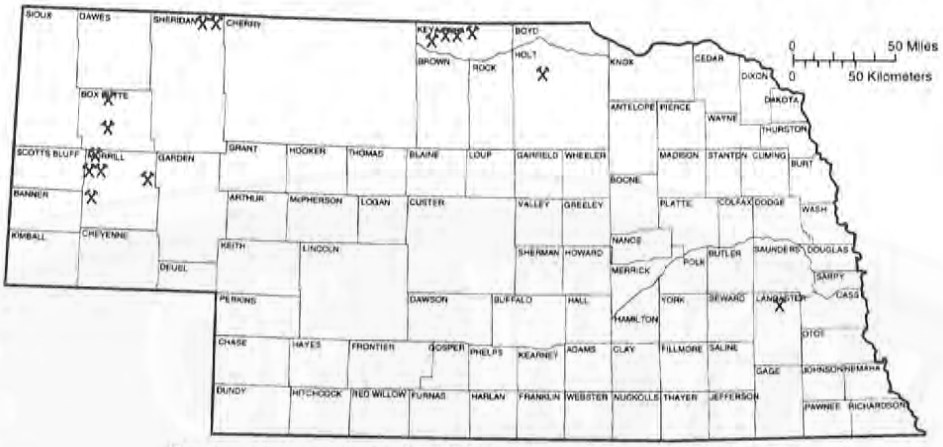


Figure 22. Locations of active sandstone pits in Nebraska.

Uranium

Uranium is a silvery white metal consisting of three radioactive isotopes U^{238}, U^{235}, and U^{234}. It occurs in more than 50 minerals, most of which are rare. The most common minerals are uraninite and other secondary minerals formed by weathering such as carnotite and autunite.

Prior to 1942, uranium was used primarily for making ceramic glazes and coloring glass. From 1945 to the late 1980s the primary use of uranium was in nuclear explosives. Since that time uranium has become an important source of energy.

Minor amounts of uranium have been known in Nebraska since the 50's, but it was not until 1980 that a world-class deposit was found near Crawford in Dawes County of northwestern Nebraska (Burchett, 1990, revised 1996; **Figure 23**). The uranium ore occurs at depths of 400 to 800 feet below land surface in the Chadron Sandstone of Tertiary age. In the future other uranium deposits may be found in Tertiary age sands or sandstones in western and central Nebraska. Sandstones of Cretaceous age in eastern Nebraska also could contain uranium (Burchett, 1990m).

Estimated uranium production in Nebraska for 1995 was 729,000 pounds of yellowcake (U_3O_8) (**Figure 24**). Its value was $8,383,500 (**Figure 25**). As of January 1, 1996, the cumulative value of uranium produced in the state was $28,306,500.

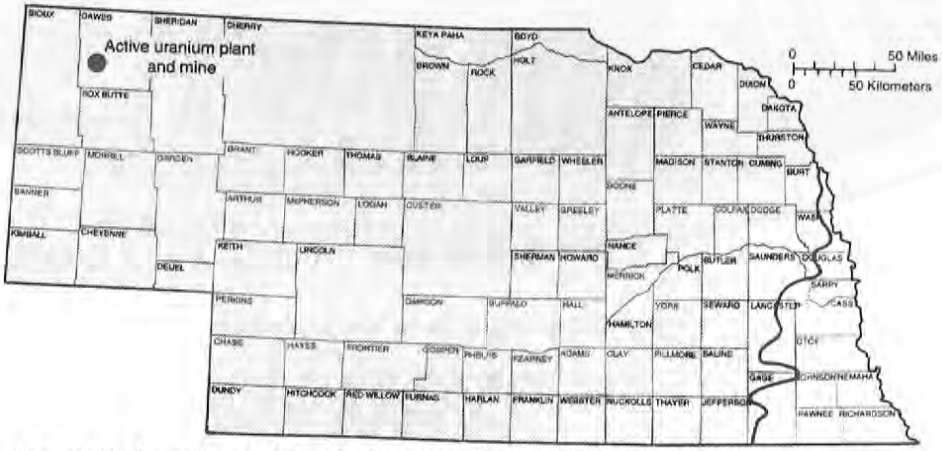


Figure 23. Location of the active uranium plant and areas of potential uranium occurrences (stippled area) in Nebraska.

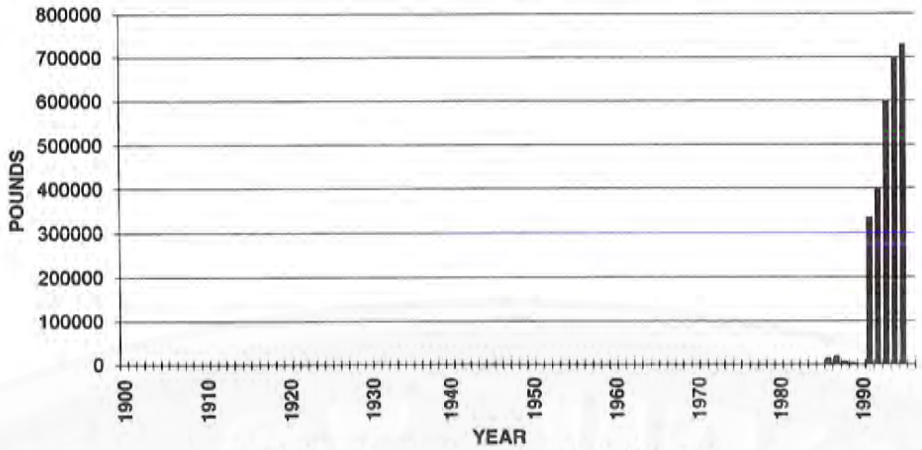


Figure 24. Annual uranium production in Nebraska.

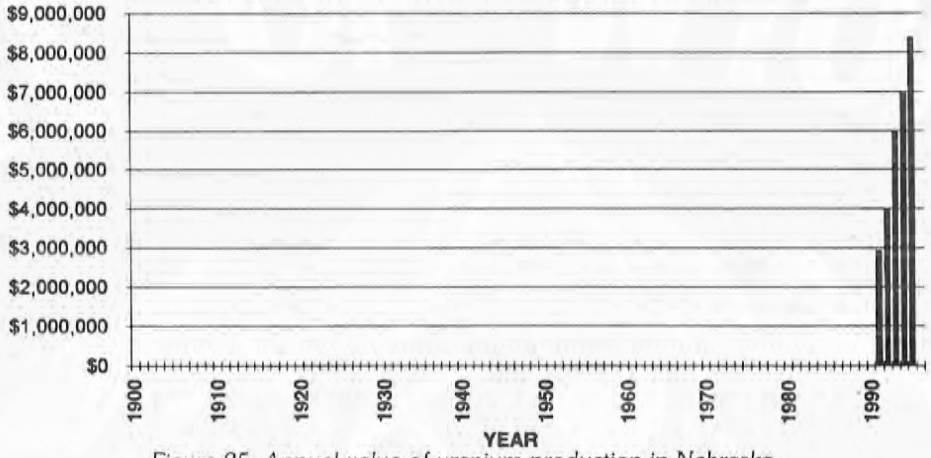


Figure 25. Annual value of uranium production in Nebraska.

Other industrial minerals

Many industrial minerals produced in the past but not currently include: chalk, flint (Burchett, 1990e), ocher (Burchett, 1991b), potash (Burchett, 1990h), quartzite (Burchett, 1990i), and volcanic ash (Burchett, 1990n). However, some of these minerals may be produced again in the future. Several other minerals known to occur in Nebraska include: barite (Burchett, 1990a), bentonite (Burchett, 1990b), diatomaceous earth (Burchett, 1990d), gypsum (Burchett, 1990f), salt (Burchett, 1990j), and zeolite.

Value of industrial minerals

The value of industrial mineral production in Nebraska during 1995 was $150 million (**Figure 26**). A comparison with the value of minerals produced in 1994 is shown in **Figure 27**. About 1500 people are

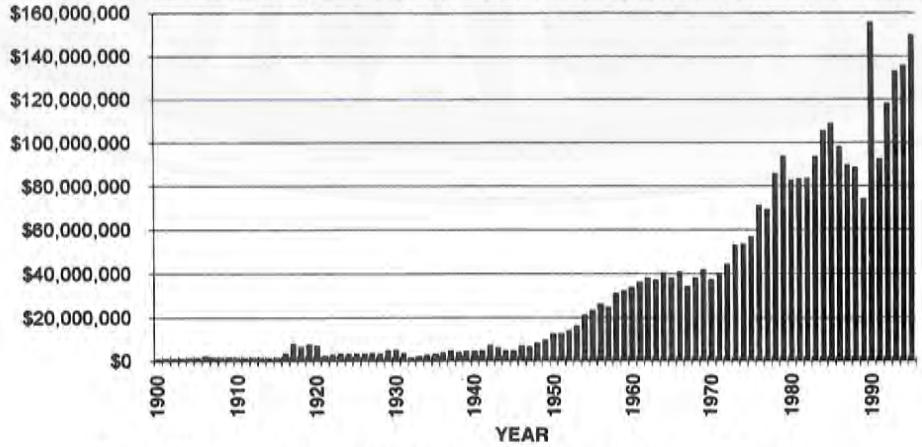


Figure 26. Annual value of industrial mineral production in Nebraska.

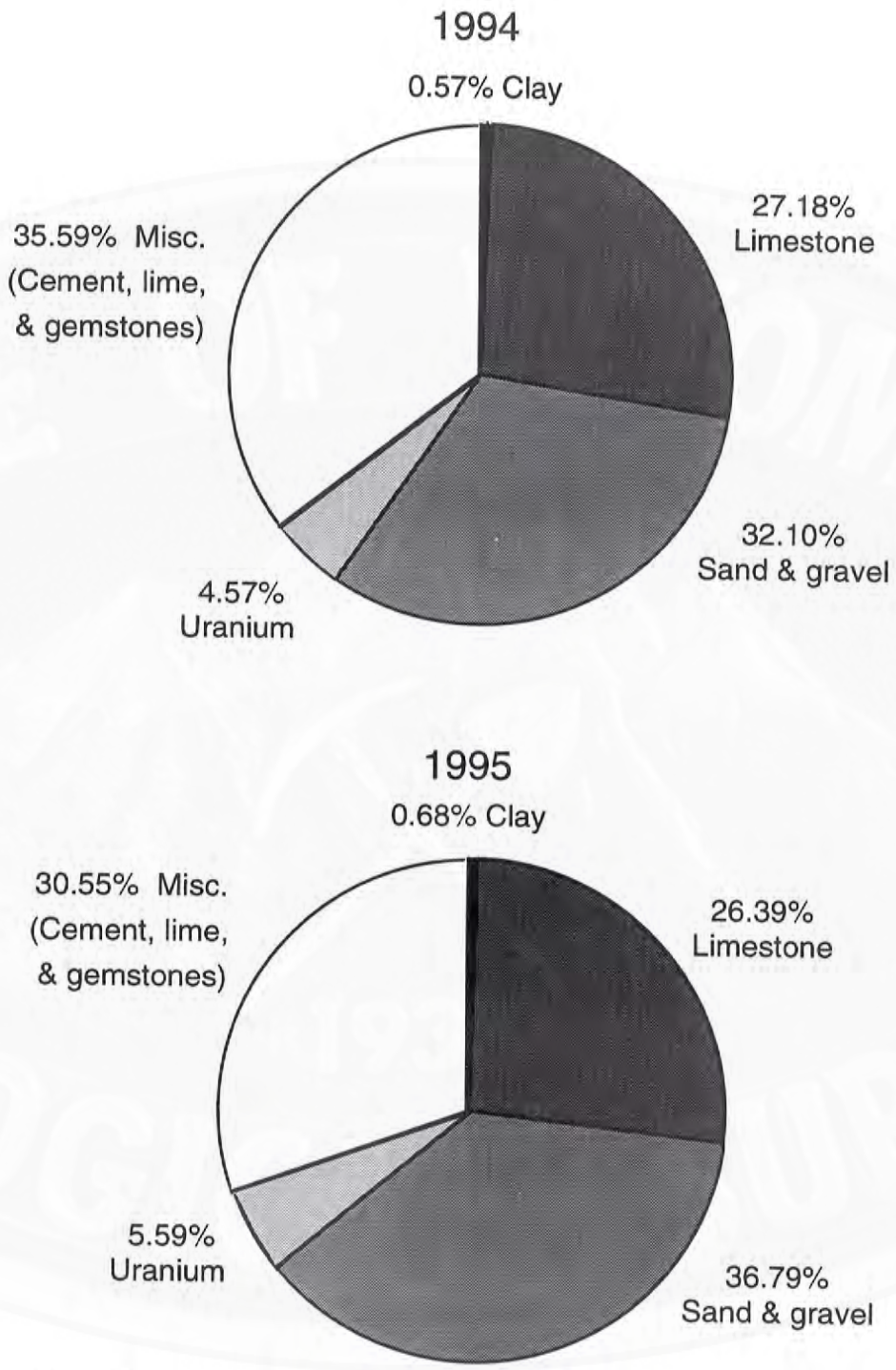


Figure 27. Relative values of Nebraska's industrial mineral production in 1994 and preliminary values of industrial mineral production in 1995.

employed by the mining industry in Nebraska, the greater number working in those counties having the largest mineral production value. The 1995 industrial mineral production of Nebraska, as reported by the U.S. Bureau of Mines and the Nebraska Oil and Gas Conservation Commission, is given in **Table 1**.

Table 1. Preliminary data on mineral production in Nebraska during 1995. Production as measured by mine shipments, sales, or marketable production (including consumption by producers). Source of data: U.S. Bureau of Mines.

Industrial mineral	Quantity[1]	Value in thousands of dollars
Clays	243	1025
Gemstones	W[2]	W[2]
Lime		668
Sand and gravel:		
Construction	17,637	55,200
Industrial	W	W
Stone (crushed)	7275	39,600
Combined value of cement, gemstones, and values indicated by W		45,172
Uranium (yellowcake)	729[3]	8384[4]
TOTAL		150,049

[1] In thousands of short tons, except for uranium.

[2] Withheld to avoid disclosing company proprietary data; value included with "Combined value" figure.

[3] In thousands of pounds.

[4] Estimated from spot price of $11.50 per pound.

Mineral resource industries in Nebraska

Industrial plants manufacturing products from mineral resources are scattered throughout Nebraska. These plants, locations of which are shown in **Figure 28**, are discussed as follows:

Agricultural lime plants

Two firms produce agricultural lime exclusively. They are located near Garland in Seward County and near Nelson in Nuckolls County. Most crushed limestone plants, largely in southeastern Nebraska, produce some agricultural lime.

Brick plants

Three firms produce brick: Endicott Clay Products Company near Endicott in Jefferson County; Yankee Hill Brick Manufacturing near Lincoln in Lancaster County; and Omaha Brick Works near Ralston in Douglas County (**Figure 29**). Common and face brick are the main products of these plants.

Cement plant

Cement is produced by Ash Grove Cement Company near Louisville in Cass County (**Figure 30**). Limestone, shale, and gypsum are the chief raw materials used in cement manufacture.

Concrete products plants

Many plants across the state produce concrete products such as block, brick, precast and prestressed slabs and construction beams, tile, pipe, vaults, septic tanks, steps, feed bunks, and bunker silos.

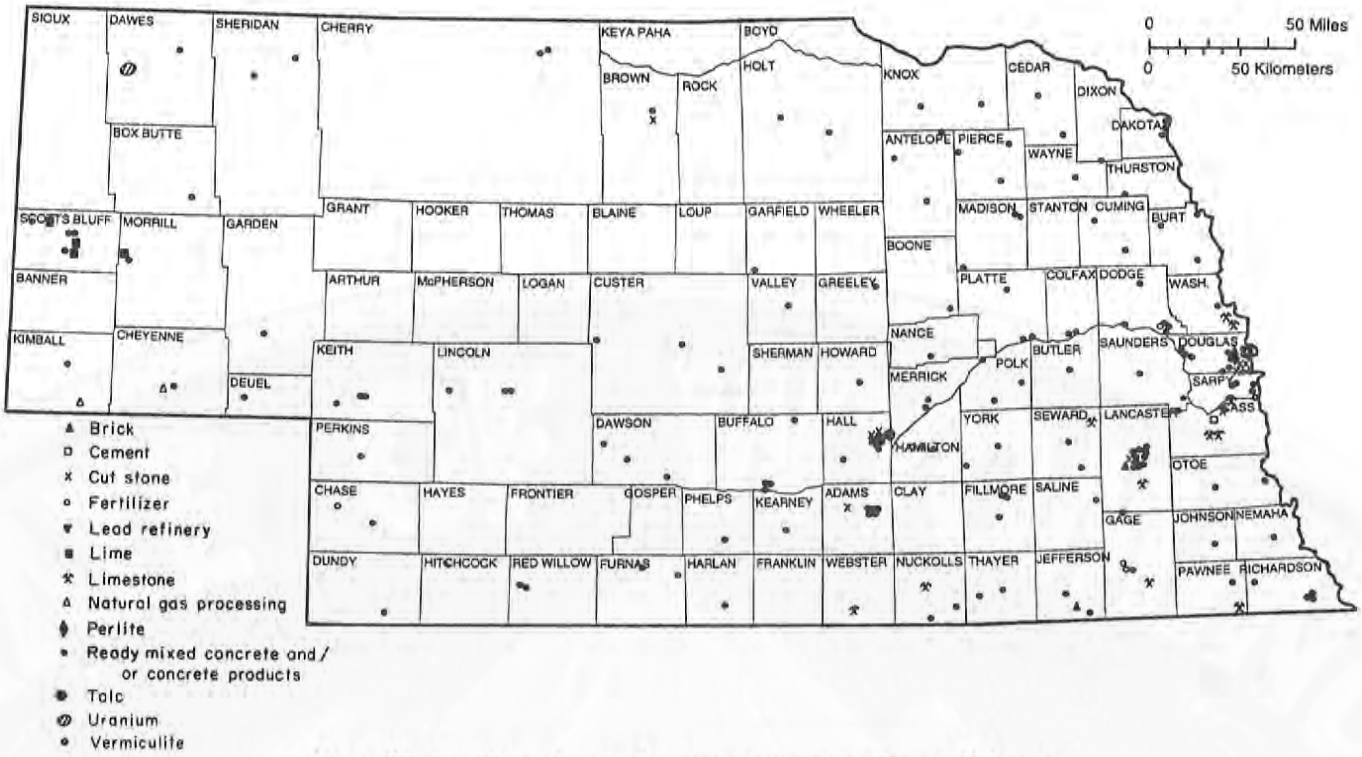


Figure 28. Locations of mineral-resource processing plants in Nebraska.

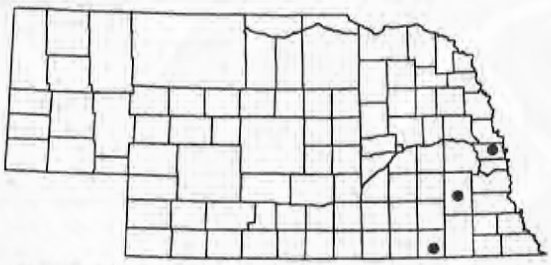


Figure 29. Locations of active brick plants in Nebraska during 1995.

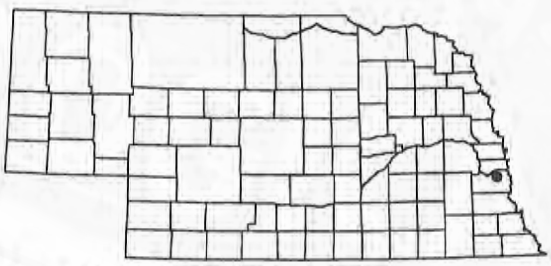


Figure 30. Location of the only active cement plant in Nebraska during 1995.

Crushed limestone plants

About 13 limestone plants are located in the eastern third of Nebraska. Principal producers are Fort Calhoun Stone Company, Kerford Limestone Company, and Martin Marietta Aggregates. Cass and Washington Counties are the leaders in production (**Figure 31**). Crushed limestone is used for aggregate in concrete, cement manufacture, road base, riprap, agricultural lime, wallstone, and mineral fillers.

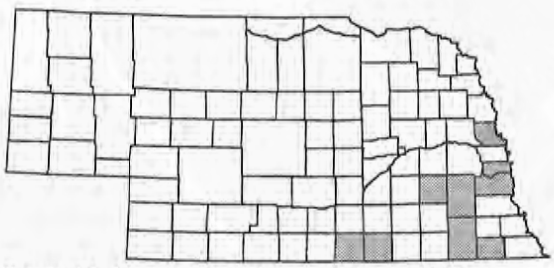


Figure 31. Locations of nine counties in which limestone was produced at 13 plants in Nebraska during 1995.

Cut-stone plants

Stone imported from other states is cut by Nebraska firms. Two of these firms are near Grand Island in Hall County, two are near Lincoln in Lancaster County, one is near Juniata in Adams County, one is near Ainsworth in Brown County, and four are near Omaha in Douglas County.

Gem and ornamental-stone plants

Many small firms around Nebraska cut and polish gems and ornamental stones such as agate, quartz, jasper, chalcedony, chert, and petrified wood.

Glass plants

Several firms around the state manufacture items from glass produced in other states.

Ground limestone plants

Three firms in Nebraska produce finely ground limestone (calcium carbonate) for feed supplements and fillers for cement, paint, and rubber. Kerford Limestone Company and Iowa Limestone Company are west of Weeping Water in Cass County, and Texas Gulf is southeast of Weeping Water.

Lime plants

Western Sugar Company produces lime from limestone imported from their quarry in Wyoming. Kilns are located at Scottsbluff and Mitchell in Scotts Bluff County, and at Bayard in Morrill County (**Figure 32**). The lime is used mainly for sugar refining.



Figure 32. Locations of active lime kilns in Nebraska during 1995.

Limestone pelletizing plant

Ampel Corporation of Des Moines, Iowa makes limestone pellets. Their plant is located on a reclaimed quarry site west of Weeping Water in Cass County (**Figure 33**). The patented pellets are for agricultural, lawn, and garden use.



Figure 33. Location of the only active limestone pelletizing plant in Nebraska during 1995.

Perlite plant

The Zonolite Division of W. R. Grace and Company is the sole manufacturer of expanded perlite in Nebraska. Its plant near Omaha in Douglas County imports crude perlite from other states, expands it, and sells it as filler material, as aggregate for plaster and concrete, and as a horticultural product.

Ready-mix concrete plants

Ready-mix concrete plants located throughout the state produce wet concrete for construction purposes. Some of these plants also manufacture cured concrete products.

Sand and gravel plants

All but eight of Nebraska's counties produce sand and gravel (**Figure 34**). Among the largest producers are Central Sand and Gravel Company in Butler, Madison, Pierce, and Platte Counties; Lyman-Richey Sand and Gravel Corporation in Cass, Dodge, Douglas, Morrill, Platte, Sarpy, and Saunders Counties; Martin Marietta Materials in Douglas and Sarpy Counties; and Western Sand and Gravel Company in Cass, Dodge, and Saunders Counties.

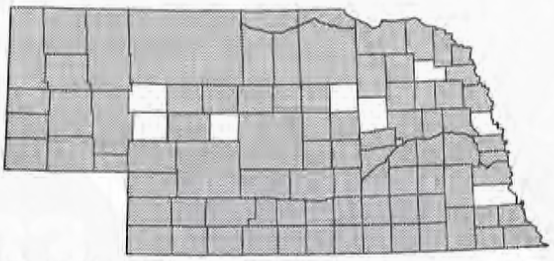


Figure 34. Nebraska counties in which sand, gravel, and silt were produced in 1995.

Talc plants

Cyprus Mines Corporation (United Sierra Division's plant near Grand Island in Hall County) is the only producer of ground talc. The unground talc is obtained from outside the state. The product is used in the manufacture of paper, ceramics, rubber, paint, insecticides, textiles, and toilet articles.

Uranium plant

Crow Butte Resources, Inc., produces ore known as yellowcake at their plant and *in-situ* (in place) mine located southeast of Crawford in Dawes County (**Figure 35**).

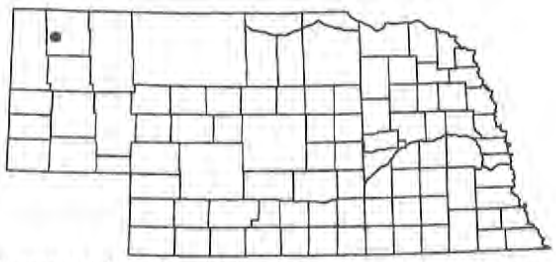


Figure 35. Location of the only active uranium plant in Nebraska during 1995.

Vermiculite plant

The only producer of exfoliated vermiculite is the Construction Products Division of W.R. Grace & Company near Omaha in Douglas County. Montana is the source of the crude vermiculite. The expanded product is used principally for insulation, concrete aggregate, and fireproofing.

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Interstate 80 crosses the Laramie Mountains between Cheyenne and Laramie and passes through part of the Medicine Bow National Forest. View to south from near summit of old Highway 30. Precambrian rocks crop out along the top of the Laramie Mountains, the Pennsylvanian Fountain Formation is found in the tree-covered slope to the right of the highway, and white limestone beds in the Pennsylvanian and Permian Casper Formation form the ridge at the top of the slope on the right side of the photograph. The Colorado Front Range forms the skyline in the upper right part of the photograph. (Photograph courtesy of Wyoming Department of Transportation.)

Industrial minerals of Colorado – review and update

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Abstract

Colorado currently produces a variety of industrial minerals and construction materials ranging from peat to diamonds. Peat, which is used as an organic soil supplement, is produced in relatively minor amounts (<2200 short tons in 1995) from alpine bogs.

Clay minerals are mined predominantly from Cretaceous Dakota Group sedimentary rocks along the Front Range where the centers of population and industry are located. The principal use is in brick fabrication although some kaolin clays are used for ceramics.

Crushed stone and sand and gravel accounted for $157 million in produced material in 1995, about a third of the state's total non-fuel mineral production. As easily accessible sand and gravel deposits are depleted along the Front Range, more producers are turning towards crushed stone. Political and environmental concerns are the dominant factors in the permitting of new crushed stone mines.

Scoria is mined as a lightweight aggregate in Costilla and Eagle Counties. The source material at the quarry in Eagle County is the state's youngest volcanic rock, the Dotsero flow, dated at 4000 years before present (y.b.p.). Other lightweight aggregates are produced from the Pierre Shale in Jefferson County.

Gypsum is mined from the Pennsylvanian Eagle Valley Evaporite by open pit methods in Eagle County where it is processed into wallboard and other products. Food-grade sodium bicarbonate is produced by a solution mining process in that part of Rio Blanco County in the Piceance Basin. The producing horizon, the Boies Bed of the Tertiary Green River Formation, is nahcolite (sodium bicarbonate hydrate) with minor amounts of shale.

White marble from the Mississippian Leadville Limestone is produced at the Yule Marble quarry in Gunnison County. This marble is mined in 15- to 20-ton blocks and is shipped to various fabricators in the United States and abroad.

Production of gem and industrial quality diamonds commenced in May, 1996 at the Kelsey Lake mine in the State Line district of Larimer County. The largest diamond found to date is a 14.2-carat, gem quality, white diamond. Diamonds have been known since the mid-1970s in the district; however, the Kelsey Lake mine will be the district's first commercial enterprise.

Industrial mineral deposits that may soon be developed include the perovskite (calcium titanium oxide) resource of the Powderhorn carbonatite, Gunnison County. Titanium oxide is used primarily as a whitener in the paint and pigment industry. Other industrial mineral commodities such as alunite, beryl, dolomite, feldspar, limestone, mica, perlite, and salt occur in Colorado but remain uneconomic because of their grade, remote occurrence, or low commodity price.

Introduction

The earliest recorded exploitation of Colorado's industrial mineral resources is the mining of Dakota Group clay and clayrock along the Colorado Front Range in the 1860s. These materials were used in the manufacture of bricks for the growing communities of Denver and Golden and in the gold fields of Central City and Blackhawk (Waagé, 1961).

Today many types of commodities are produced from industrial mineral deposits in Colorado, ranging

from peat in Holocene alpine peat bogs to diamonds in Devonian kimberlites of the State Line district. Construction materials such as clay, limestone, sand, gravel, crushed stone, and gypsum are an important part of Colorado's mining economy and their production is a reflection of the state's economic health. Sand and gravel and crushed stone alone accounted for $157 million worth of material, about a third of the total value of Colorado's non-fuel minerals produced in 1995. **Figure 1** shows selected industrial and gem mineral occurrences in Colorado that will be mentioned in this report.

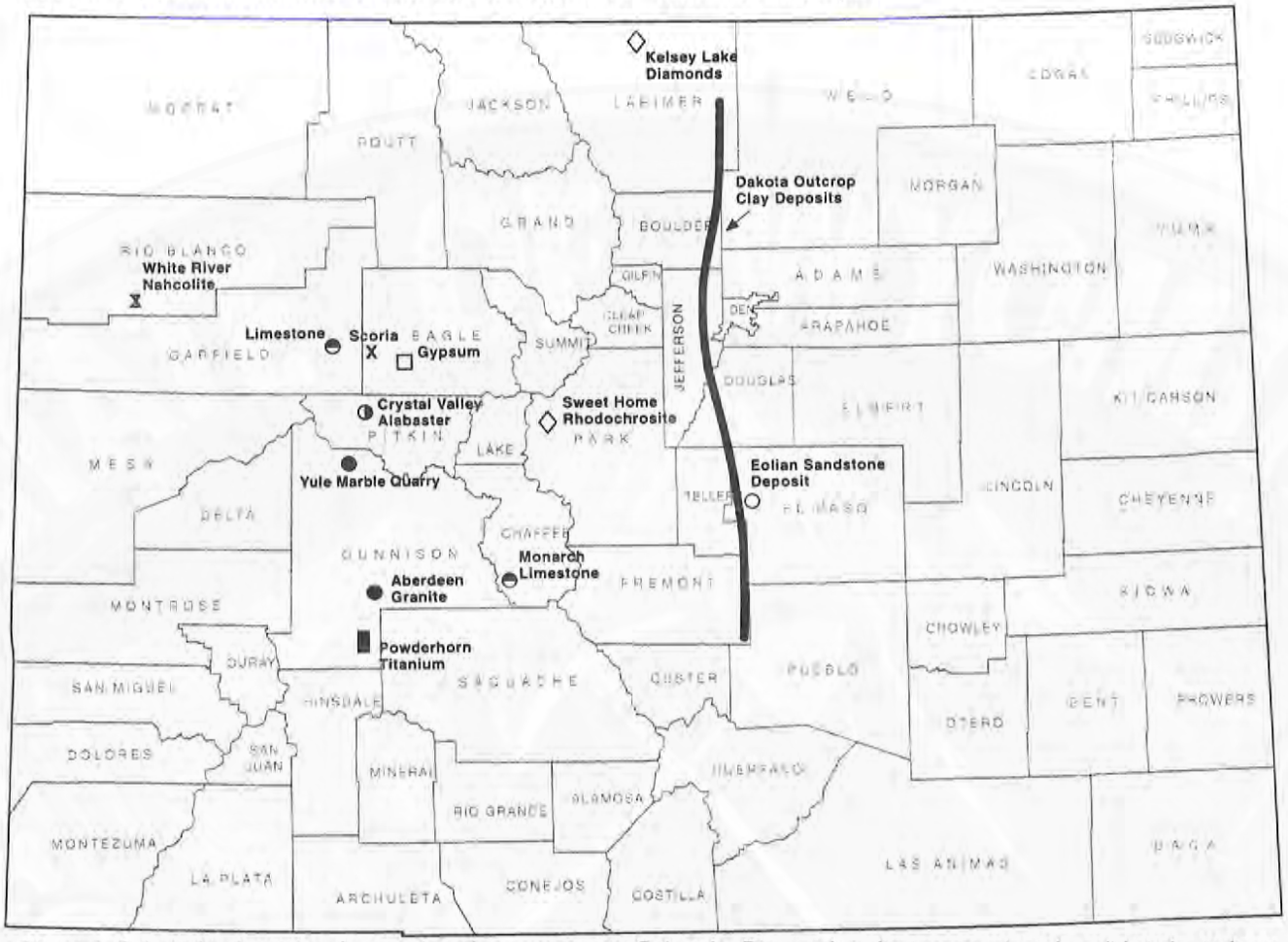


Figure 1. Selected industrial and gem mineral occurrences in Colorado. Diamonds indicate gem minerals, solid circles indicate dimension stone quarries.

Peat

Peat is a naturally-occurring organic-rich substance derived primarily from the preservation of plant materials. Peat is used primarily as a plant-growth medium and soil supplement by the horticultural and agricultural sectors. In Colorado, peat is mined in small amounts from alpine bogs in about four locations. The total amount of peat produced in 1995 was small, less than 2200 short tons.

Sand, gravel, and aggregate

Sand, gravel, and aggregate production peaked in 1993 (**Figure 2**) with the construction of several major civic projects such as Denver International Airport and Coors Field, home of the Colorado Rockies baseball team. Production in 1995 is estimated at 40 million short tons. New home construction and road repair projects account for much of the new sand, gravel, and aggregate production in the state.

Sand and gravel is becoming less available along the Colorado Front Range because of two main factors–depleted reserves and lack of available land for quarries. In other parts of the state such as along the Eagle River in Eagle County, a region that has experienced a significant population growth in recent years, sand and gravel resources are plentiful and are still being quarried (**Figure 3**).

Much of the construction material along the Front Range comes from aggregate quarries in Proterozoic metamorphic rocks and Tertiary volcanic rocks. As political pressures mount for open space and residential development, it is becoming more difficult to permit mountain and foothill land for quarries.

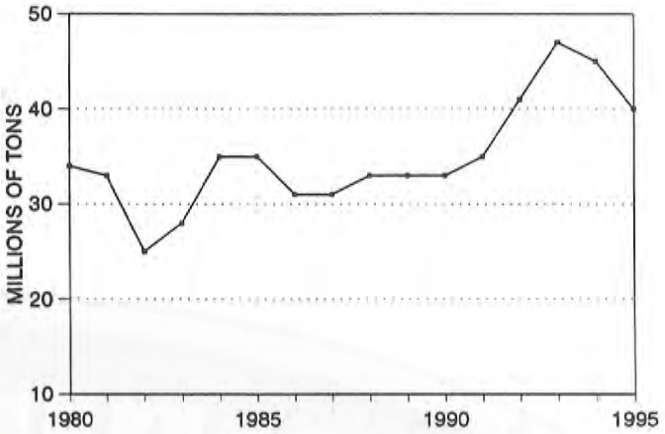


Figure 2. Production of sand, gravel, and aggregate in Colorado from 1980 to 1995.



Figure 3. Sand and gravel quarry along the Eagle River in Eagle County. Note mobile crushing and screening plant.

Clay

Clay production in Colorado has been steady at about 300,000 short tons over the past three years (**Figure 4**). Most of the clay minerals mined in Colorado are either bentonites or refractory clays. The most important bentonite deposits are alteration products of Tertiary rhyolitic tuffs and occur near the historic silver mining town of Creede. Other deposits occur in the Silver Cliff and Rosita Hills district and near Salida; the deposits occur in altered Tertiary volcanic rocks (Patterson, 1964).

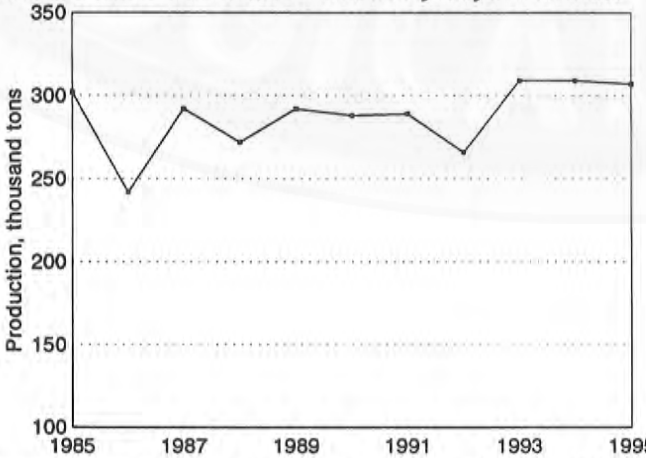


Figure 4. Production of clay in Colorado from 1985 to 1995.

Refractory clay deposits occur primarily in the Cretaceous Purgatorie Formation and Dakota Group and in the Tertiary Dawson Arkose. These formations outcrop along a narrow, moderately-dipping belt or in widespread areas of essentially horizontal dip along the Front and Wet Mountain Ranges of Colorado (**Figure 1**).

The location and physical characteristics (including Pyrometric Cone Equivalents) of most Colorado clay deposits have been well described by various authors including Butler (1914), Argall (1949, p. 89-109), Van Sant (1959), and Patterson (1964).

The Dakota Group has been divided into several formations along the Front and Wet Ranges (Waagé, 1953, 1961). The principal source of refractory clay in this group is the Van Bibber shale member which is composed of alternating sandstone and shale units. The shale units generally consist of a basal kaolinitic layer, overlain by a thin zone of silty clay shale or siltstone. Another important clay-producing unit is the so-called ripple clay shale, a source of high refractory clay in the northern Front Range; it is a relatively silt free, finely laminated, dark gray to black, clay shale (**Figure 5**) (Waagé, 1961, p. 30). In the southern Front Range the most valuable clay deposits occur in the blue-gray, plastic and flinty, clay shales of the Dry Creek Canyon Member of the Dakota Group (Waagé, 1953, p. 14).



Figure 5. Parfet Clay's quarry in Dakota Group rocks south of Golden, Jefferson County.

The Purgatorie Formation, which occurs in south-central and southeastern Colorado, is now recognized as part of the lower Dakota Group (T. Hemborg, 1996, oral communication). The Glencairn shale member consists of alternating units of olive brown sandstone and gray and black shale. The shale consists of generally low refractory, plastic, blue-gray clay (Waagé, 1953, p. 30). Most clay produced in Colorado goes into the manufacture of bricks and other masonry products. Some of the kaolins are used in specialized ceramic applications.

Lightweight aggregate

Lightweight aggregates are specialty products for the construction industry where weight is an important factor. Rock materials suitable for lightweight aggregate include vermiculite, perlite, scoria, clay, and shale. The Cretaceous Pierre Shale, which has a favorable clay mineralogy (dominantly montmorillonite or illite) and low amounts of sand- and silt-sized material, is mined by Western Aggregates, Inc. at a site in northern Jefferson County. The shale material, which has an in-place density of about 80 to 90 lbs/ft^3(cubic feet), is crushed and heated in a rotary kiln at a temperature of 2000°F. The final processed material has a bulk density of 40 lbs/ft^3. Western Aggregates produces about 300,000 yd^3 (cubic yards) of lightweight aggregate per year. Their main markets are concrete roof tiles (mainly for the California market), chip seal aggregate for road bases in the Rocky Mountain region, and concrete blocks.

The Mayne Block Company at Dotsero, in Eagle County, produces about 10,000 yd^3 per year of lightweight aggregate from volcanic scoria at a small quarry in the agglomerate beds of the 4150-year-old Dotsero volcanic complex (Giegengack, 1962) (**Figure 6**). The material is screened and crushed at the mine and plant for use in concrete blocks or as chip seal for road base. The central Colorado region is the main market area for this plant.



Figure 6. Scoria quarry in the Dotsero volcanic complex near Dotsero in Eagle County. Note crushing and screening plant in lower part of the photograph.

Dimension stone

The Yule Marble quarry at Marble, in Gunnison County reopened in 1990 after a five-decade hiatus. Yule Marble, known for its nearly pure white color, has been used in many famous buildings including the Tomb of the Unknown Soldier and the Lincoln Memorial in Washington, D.C., as well as in parts of the

Colorado State Capitol, Post Office, and Customs House in Denver. The marble is quarried from a vast underground stope and cut into 20-ton blocks by special chain saws (**Figure 7**). The blocks are transported out of the quarry (**Figure 8**) by truck to the nearest railhead in Glenwood Springs where they are sent to fabricators in the United States and overseas.



Figure 7. Yule Marble's quarry at Marble, Gunnison County. Note front end loader at left of photograph for scale.



Figure 8. Cut and trimmed blocks of Yule Marble partially shown at left of photograph. Photograph taken just outside the entrance to the quarry. Blocks are awaiting shipment to the railhead at Glenwood Springs.

In 1995 the company produced 4400 short tons of salable marble. Approximately 20% of the production was sent overseas, primarily to Italy, Japan, and Indonesia, where it is fabricated into consumer products. The demand for stone floor tiles within the United States is increasing as the price disparity between ceramic and stone tiles diminishes. The Colorado Yule Marble Company has purchased a new production saw and plans to add a third shift in 1996 in order to approximately double production to satisfy the increased demand for high quality marble tiles.

The Yule Marble was formed by recrystallization of the Mississippian Leadville Limestone during the intrusion of the Tertiary quartz monzonite porphyry of the Raspberry Creek phacolith (Gaskill and Godwin, 1966).

The Crystal Valley Mining Company is conducting limited development work on their black alabaster and marble deposit in Pitkin County. The deposit is hosted by the Pennsylvanian Belden Formation which was recrystallized by intrusion of the Tertiary Mount Sopris quartz monzonite stock. The black marble is noted for its hardness and multicolored veinlets. The project, which is located on U.S. Forest Service land, has received the necessary permits from state and Federal authorities, but Crystal Valley is still in the process of obtaining a permit from Pitkin County.

There are several other deposits of marble in Colorado that are no longer in production. In Chaffee County, a white marble is found at the Monarch mine. Colored marbles are found in Chaffee, Fremont, Pitkin, Pueblo, Saguache, and Routt Counties (Argall, 1949, p. 285-287).

Several occurrences of granite in Colorado have been used over the years as dimension stone and still have potential for development. The Aberdeen quarry in Gunnison County produced a fine-grained, gray granite that was used in the construction of the State Capitol and the old Colorado State Museum. Granite from Cotopaxi in Fremont County was used in the construction of the Denver City and County Building, the Denver Mint, and other buildings in Denver. Granite is currently being quarried in small amounts from locations near Grand Junction and Estes Park.

There are several other dimension stone quarries in Colorado that are currently active or semi-active. The Wall Mountain Tuff, a pink rhyolite tuff with few accessory minerals, is quarried near Castle Rock in Douglas County. Quarries for red sandstone and flaggy red sandstone, as well as lighter-colored sandstone, continue to operate at several localities throughout the state. The tuff and the sandstones can be used for dimension stone once the market improves and production costs go down.

Specialty sand

Each year Colorado Silica Sand, Inc. of Colorado Springs produces approximately 200,000 short tons of specialty sand that is used for fracturing in oil and gas wells, filter media for water purification plants, gravel packs around water wells, and other applications where roundness, permeability, and strength are important parameters. About 50% of the production is shipped overseas, primarily to Canada, South America, and Europe.



Figure 9. Outcrop of Quaternary eolian sandstone along the Old Ranch Road in El Paso County. Note cross-bedding. Camera lens in middle of photograph for scale.

The sand comes from several quarries in Quaternary eolian deposits located around the northern Colorado Springs area (**Figure 9**). The sand in these deposits is fine- to coarse-grained, well rounded, and has a thickness of about 20 feet (Scott and Wobus, 1973). The current procedure is to quarry these eolian sands and then recontour and reclaim the surface prior to residential development. The raw sand is trucked to the plant where it is washed, classified, dried, and screened to produce several mesh-size products for different markets (**Figure 10**).

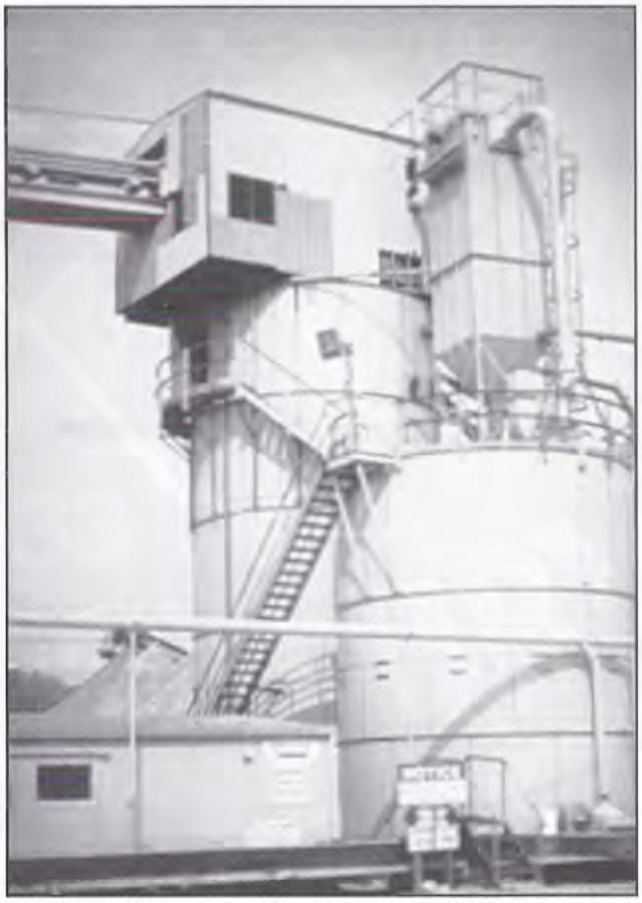


Figure 10. Screening tower at Colorado Silica Sand, Inc.'s facility, southern Colorado Springs, El Paso County.

Gypsum

Gypsum is found in abundance throughout Colorado in sedimentary rocks mostly of Pennsylvanian and Permian age. The most notable occurrences are in the Eagle Valley Evaporite of the Eagle Basin in Garfield and Eagle Counties and in the Paradox Formation of the Paradox Basin in southwestern Colorado. Other gypsum occurrences of potential economic importance are located in the Lykins and Morrison Formations in the Front Range and in the Minturn Formation of Chaffee, Park, and Fremont Counties (Withington, 1964).

Gypsum is used primarily in the fabrication of wallboard and for construction plaster. Uncalcined gypsum is used as a retardant in the cement industry and as an agricultural mineral (soil conditioner).

Gypsum production in 1995 at the Eagle gypsum mine in Eagle County was at the nameplate capacity of 400,000 short tons, the same as 1994. The gypsum ore is mined from massive gypsum beds of the Eagle Valley Evaporite in an open pit mine. Pavement profiler machines cut a trench 12 feet wide by 0.5-foot deep (**Figures 11** and **12**). The gypsum is manufactured into wallboard and other products at the plant in Gypsum. Approximately 50% of the wallboard pro-

duced at the plant goes to the Colorado construction industry; the remainder is marketed throughout the United States.



Figure 11. Eagle Gypsum's open pit mine just north of the town of Gypsum in Eagle County. Eagle River valley in the background. Pavement profiler machines cut a trench and pile the broken gypsum ore into furrows and a front end loader moves the broken ore to a truck loading site.



Figure 12. Pavement profiler at the Eagle Gypsum mine. Note furrow of gypsum ore at the rear of the machine.

Sodium bicarbonate and other alkali minerals

Nahcolite ($NaHCO_3$) and halite (NaCl) occur as aggregates and in disseminated and massive beds with oil shale in a saline facies of the Parachute Creek Member of the Eocene Green River Formation in the Piceance Basin, northwestern Colorado. Nahcolite resources are estimated to be as high as 489 million tons per square mile, and the total *in situ* resource is estimated at 32 billion tons (Dyni, 1974). Sodium bicarbonate, the chemical produced from nahcolite, is used mostly as a food additive (in the form of baking soda), in various household and industrial chemicals, and for other miscellaneous uses including the removal of toxic compounds from flue gas.



Figure 13. The White River Nahcolite Company plant in Rio Blanco County. The beehive-shaped structure on the right of the photograph is a storage facility for sodium bicarbonate.

During 1995 White River Nahcolite Company's nahcolite solution mine and plant in Rio Blanco County produced approximately 50% of its nameplate capacity of 125,000 tons a year (**Figure 13**). The company produces both food and industrial grade sodium bicarbonate from the 2000-foot-deep, 20- to 25-foot-thick "Boies Bed," one of many nahcolite-bearing beds in the Parachute Creek Member. The company is currently producing from solution cavity number 4 (**Figure 14**) which has a projected yield of 125,000 tons at a well spacing of 3000 feet. Recent permit changes have allowed the company to reopen



Figure 14. Vertical recovery well, solution cavity number 4, White River Nahcolite Company, Rio Blanco County.

solution mining operations at solution cavity number 1. The solution cavities as formed have a different shape, which allows for greater recoveries, than early design plans indicated.

In the Paradox Basin, layers of halite and potassium salts are found interbedded with gypsum and anhydrite and clastic rocks in the Pennsylvanian Paradox Formation. Halite is also known to occur in the Eagle Basin where an oil well test hole near Glenwood Springs terminated after intercepting 935 feet of halite and other salts in the Eagle Valley Evaporite (R. Kirkham, 1996, oral communication).

Halite also occurs in the Permian rocks of the Denver Basin in northeastern Colorado. Permian rocks which outcrop along the Front Range are coarse clastic deposits; however, to the east deep oil well tests have intercepted several halite-bearing beds and carbonate rocks in the Permian strata (Raup, 1964).

Titanium

Titanium oxide minerals occur in three principal localities in Colorado. Titaniferous magnetite occurs within the small pyroxenite intrusions of the Caribou district in Boulder County. These deposits tend to be comparatively low grade and of small size. Titaniferous magnetite occurs in gabbro and pyroxenite dikes in a region southwest of Cañon City. The dikes range between 10 to 30 feet in width and about 1600 feet in length (Argall, 1949, p. 455-456). The most significant resource of titanium oxide in Colorado is located within the Powderhorn carbonatite in southern Gunnison County.

Titanium dioxide is used primarily as a pigment for surface coatings in the paint and paper industries. Titanium dioxide has a high refractive index which imparts whiteness, opacity, and brightness to paints and paper. Titanium pigment is also used in the plastics industry where it serves to prevent the degradation of plastic products. Titanium metal, because of its light weight and strength, is used primarily in the aircraft and aerospace industries.

The Powderhorn district is of interest because it contains the largest known reserves of both titanium and niobium in the United States. The district is also enriched in rare-earth minerals, thorium, iron, uranium and vanadium, and contains economic deposits of vermiculite. Early economic interest in the district focused primarily on the vermiculite resources of the area and small amounts were produced in the 1930s and 1940s (Temple and Grogan, 1965). The large resource of titanium in the Powderhorn district, contained in the mineral perovskite, was first described in 1957; however, attempts at development have not been successful due to economic reasons. The niobium and rare-earth resources have also not been exploited to date.

The Powderhorn district is located in an area of Precambrian basement rocks on the northwestern side of the San Juan volcanic field. Larsen (1942) published the first comprehensive report on the geology, geochemistry, petrology, and mineralogy of the Iron Hill complex. Temple and Grogan (1965) published a detailed account of the geology and economic potential of the district.

The Powderhorn carbonatite complex was intruded into the surrounding Proterozoic igneous rocks as a pear-shaped mass of carbonatite and other alkalic rocks. The alkalic rocks comprising the complex, from oldest to youngest, are pyroxenite, uncompahgrite, ijolite, nepheline syenite, and carbonatite. Potassium-argon age determinations from a number of minerals give an age of emplacement for the complex at 570 Ma (Mega-annum or million years before present) (Olson and others, 1977). The titanium resource at Powderhorn is believed to be the largest in the United States, comprising over half of the known usable reserves of titaniferous raw material (Thompson, 1987). The main ore mineral is perovskite, a calcium-titanium oxide. Titanium reserves have been estimated at 390 million short tons of ore assaying 11.5% TiO_2, according to an unpublished report by Kaiser Engineers written in 1976.

Several titanium-bearing minerals are present in the pyroxenite of the Iron Hill complex including

perovskite, ilmenite, leucoxene, sphene, melanite, garnet, and titaniferous magnetite. Minor amounts of titanium occur in diopside and biotite.

The major problem confronting development of the titanium resources at Powderhorn has been the complex processing required to treat perovskite and the other titanium-bearing minerals. Teck Resources has been conducting an evaluation program on the Powderhorn titanium resource since 1990. No work other than continued market evaluation took place in 1995. In 1996 Teck plans to conduct metallurgical testing on the ore and is likely to undertake a prefeasability study.

Limestone

Limestone has been used for a variety of purposes including the sugar beet industry, metallurgical limestone, cement, chemical grade calcium carbonate, and aggregate. The decline of the sugar beet industry and the cessation of steel production at the CF&I plant in Pueblo have reduced the demand for limestone in Colorado.

Limestone and marlstone are quarried from the Niobrara Formation of Late Cretaceous age in places in eastern Colorado. Paleozoic carbonate rocks such as the Leadville Limestone, the Ingleside Formation, and, locally, other lower Paleozoic beds have been mined and continue to be mined for limestone in Colorado.

Alunite

Alunite ($KAl_3(SO_3)_2(OH)_6$) from the Marysvale district, Utah was used as a source of potash during World War I. It was considered as a potential source of aluminum prior to and during World War II. Although Colorado has many significant occurrences of alunite, none of these has ever been exploited as a source of either potash or aluminum.

Alunite occurs as a secondary mineral primarily in areas of intense acid-sulfate alteration of felsic volcanic rocks. The principal occurrences in Colorado are: Marble Mountain and Summitville district in Rio Grande County; Rosita Hills and Silver Cliff district in Custer County; Calico Peak in Dolores County; and Red Mountain in Hinsdale County (Ratté, 1964; Hall, 1980). The Calico Peak deposit contains a reserve of approximately 6 million tons at a grade of 17% alunite (Argall, 1949, p. 20-22). A resource area defined by Hall (1980) at the Red Mountain deposit consists of 250 million tons at an estimated grade of 30% alunite.

Gemstones

Gem mining in Colorado has had a long and varied history including the production of agates, amazonite, aquamarine (the state mineral), chrysoberyl, lapis lazuli, opal, phenacite, rhodochrosite, quartz, topaz, and turquoise (Argall, 1949, p. 217-218). In 1996 diamonds and rhodochrosite mining will be the most significant gem-producing activities in Colorado.

Diamonds were discovered in kimberlites of the State Line district in 1975. About 100 kimberlites have been identified in the northern Front Range of Colorado and the geologically continuous Laramie Mountains of southern Wyoming (**Figure 15**). The oldest rocks in the region are meta-



Figure 15. Xenoliths of Proterozoic granite (dark-colored rock fragments) and Paleozoic limestone (light-colored rock fragments) in the Kelsey Lake kimberlite, Larimer County.

sedimentary gneisses of middle Proterozoic age which have been intruded by two upper Proterozoic batholiths, the Log Cabin Granite and the Sherman Granite. The deposition of lower Paleozoic clastic and carbonate sedimentary rocks is confirmed by the presence of exotic blocks of those rocks in the kimberlite intrusive diatremes of the district. After the kimberlites were intruded during the Devonian (Eggler, 1968), the entire Front Range region was uplifted, all the lower Paleozoic rocks were removed by erosion, and the upper Paleozoic redbeds of the Fountain Formation were deposited.

There are approximately 40 kimberlite diatremes and dikes in the State Line district, most of which are known to be diamondiferous. They range in size from a few feet to nearly 1800 feet in diameter and, generally, have an ellipsoidal to elongate shape. Only a few of the diatremes show a definite structural control, mostly along east-west- and northwest-trending structures. Others, such as the Schaeffer diatremes, are aligned in a northwest direction, oblique to the local fracture direction (Hausel and others, 1985).

Four main varieties of kimberlite have been recognized in the State Line district: massive to porphyritic, carbonate-rich massive to porphyritic, breccia, and carbonate-rich breccia (Smith and others, 1979). The massive to porphyritic kimberlite is best exposed in the Iron Mountain district of Wyoming and in the kimberlite dikes near Estes Park, Colorado. This type of kimberlite contains olivine and enstatite phenocrysts set in a fine-grained matrix of serpentine, calcite, dolomite, phlogopite, magnetite, perovskite, and other minerals. Serpentine, dolomite, hematite, magnetite, and iddingsite replace olivine phenocrysts. Massive to porphyritic kimberlite contains only a small percentage of lithic xenoliths.

The carbonate-rich massive to porphyritic kimberlite is texturally similar to the above described kimberlite variety; however, it contains a significantly greater amount of carbonate, most of which occurs as a replacement product of phenocrysts, xenocrysts, and serpentine in the matrix.

Kimberlite breccia and carbonate-rich kimberlite breccia contain significant numbers of these types of xenoliths: lower Paleozoic sedimentary rocks; Proterozoic crystalline rocks including granites and schists; and upper mantle nodules of spinel and garnet peridotite, garnet clinopyroxenite and websterite, dunite, eclogite, monomineralic megacrysts, and carbonatite.

In the 1970s and 1980s several mining companies began the first round of exploration in the State Line district which led to the discovery of several kimberlites; however, none of these proved to be economic at that time. In the early 1990s Colorado Diamond Company, now a subsidiary of Redaurum Red Lakes Mines Ltd. of Toronto, began a determined exploration and bulk sampling program at the Kelsey Lake kimberlite. In 1995 the company announced a 275,000-ton-per-year trial mining program. The Kelsey Lake mine will be North America's first commercial diamond producer since the turn of the century. Open pit mining at Kelsey Lake began in late 1995; the processing plant commenced operations in May, 1996. The mine and processing plant will be upgraded to 1 million tons a year production after the two-year trial mining program. The diamonds produced at the Kelsey Lake mine will be marketed as *Colorado Diamonds*™ and *Kelsey Lake Diamonds*™.

Over 60% of the diamonds recovered in previous years' bulk sampling programs are of gem quality including a 14.2-carat white diamond, the largest ever discovered in the State Line district and the eighth largest found in North America. The Kelsey Lake prospect consists of eight kimberlite pipes; however, the trial mining program will be focused on the two largest pipes which have a surface area of about 12 acres and contain about 19 million tons of ore down to 300 feet. The processing plant will be capable of producing 200 tons per hour of gravity concentrates by using only a water slurry in four rotating pans, each about 14 feet in diameter. The concentrates are mixed with water and passed over a vibrating table coated with grease. Because of their hydrophobic character, the diamonds stick to the grease as the other heavy minerals float away to the tailings pond. Initial diamond recoveries are estimated to be up to 25,000 carats per year. (See the paper by Coopersmith, this volume, for further details on this operation.)

The Sweet Home mine, an old silver prospect in Park County, was reopened as a rhodochrosite gem mine in 1991. Over the past five years the mine has produced several world class examples of cherry red rhodochrosite crystals including one thought to be the largest in the world. Some of the larger crystals set in a quartz and sulfide mineral matrix have commanded prices in the hundreds of thousands of dollars.

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STATE OF WYOMING
1933
GEOLOGICAL SURVEY

Industrial minerals of Idaho

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Abstract

Industrial minerals comprise an important part of Idaho's mining industry. In 1994, production and processing of phosphate ore in southeastern Idaho contributed $630 million to the state's $860 million mining economy, according to the U.S. Bureau of Mines, the Idaho Mining Association, and the Idaho Geological Survey. But phosphate is not the only industrial mineral mined in Idaho. Large and small operations mine and ship a diverse suite of Idaho mineral products worldwide.

Phosphate mining in Idaho started with the rich ores of the Permian Phosphoria Formation and the cheap hydroelectric power of the region. Two companies, FMC and Monsanto, now mine phosphate ore and convert it into elemental phosphorous at their plants in Pocatello and Soda Springs. With last year's closing of Rhone-Poulenc's Montana plant, Idaho is the only state producing elemental phosphorous. It is used in a variety of industrial products from soft drinks to Monsanto's popular weed killer, Roundup. J.R. Simplot and NuWest Industries (now owned by Agrium Inc.) convert phosphate into phosphoric acid which is used to make fertilizers. The value of the raw ore is increased fourfold by the in-state processing.

Production of construction sand and gravel by over 40 operators was valued at some $53 million in 1994. The value of sand and gravel now exceeds Idaho's fabled silver production. Recent population growth is starting to strain aggregate resources in urban areas, as building continues at a rapid pace. The construction boom in the western U.S. is also providing opportunities for producers of Oakley stone (a decorative quartzite) and other dimension stone producers. One Boise company, Rocktile, cuts stone tiles, some of which are shipped to international markets. Another firm, Idaho Travertine, is one of the largest stonecutting and polishing companies west of the Mississippi River.

Pumice is mined for lightweight aggregate, landscaping rock, and specialty uses, such as the high purity, low-heavy-mineral pumice powder, which Hess Pumice exports to Europe and Asia for grinding television glass. Emerald Creek Garnet mines over 27,000 tons of placer garnets a year for filters, water jets, and abrasives. It is the largest garnet producer in the U.S. Unimin Corporation mines silica sand from ancient lake bed deposits near Emmett for use in golf course sand bunkers, abrasives, and glass making. Limestone is mined by Ash Grove Cement, Chemical Lime (manufacturer of chemical grade lime), and several smaller companies. Zeolite extracted by Teague Minerals from altered Tertiary volcanics in Owyhee County has a variety of uses. Other industrial mineral deposits of the state include clay, travertine, gypsum, diatomite, perlite, scoria, and gemstones, such as jasper, opal, and garnet.

The state's diverse and rich geologic heritage has created many opportunities. Idaho's industrial minerals operations should enjoy continued success as they expand to new applications and international markets.

Introduction

Like many western states, mining played an important role in Idaho's early history. The lure of gold, and then silver and lead, brought settlers to the mountains and valleys in the 1860s. As the newcomers built houses and cities, they used local building materials including basalt and sandstone as one of the first uses of Idaho's industrial minerals. Today, industrial minerals comprise an important part of Idaho's mining industry. In 1994, the value of nonfuel mineral production in Idaho was $861 million, a sum which includes the value of processed phosphate ore (**Table 1**). The value of mined ore production of all commodities was $340 million, according to the U.S. Bureau of Mines, the Idaho Mining Association, and the Idaho Geological Survey.

Table 1. Value (in thousands of dollars) of nonfuel mineral production in Idaho, 1991 to 1994. From Idaho Geological Survey GeoNote 36. Production, except for phosphate, is based on figures provided by U.S. Bureau of Mines. Phosphate value derived from estimate by Idaho Mining Association (IMA). IMA estimate is average value (about $104/ton) of phosphoric acid and elemental phosphorus produced from raw phosphate ore and is much greater than the value of the raw ore (about $25/ton). "Other" includes antimony, cement, clay, garnet, lime, stone (dimension and crushed), vanadium, pumice, gemstones, and sand and gravel.

Commodity	1991	1992	1993	1994
Gold	$ 39,107	$ 35,241	$ 40,919	$ 39,343
Molybdenum	13,419	28,841	0	28,990
Pb+Zn+Cu	18,400	18,000	15,670	12,446
Silver	43,807	32,131	26,231	22,762
Phosphate	585,695	546,842	568,152	630,000
Other	97,518	112,194	126,653	127,171
Total	**$ 797,946**	**$ 773,249**	**$ 777,625**	**$ 860,712**

The value of Idaho's industrial mineral production in 1994 was almost $729 million, a substantial increase over the previous year's figure of $676 million (**Tables 2** and **3**). The general trend has been a slow increase in production and value with time. For 1994, $630 million of the total is attributable to the value of Idaho-produced phosphoric acid and elemental phosphorus produced from the phosphate ore. Phosphate is Idaho's leading mineral commodity, worth far more than the state's metal production, which was worth approximately $103 million. Processed phosphate is typically worth about $105/ton versus $25/ton for the raw ore, as calculated from production data collected by the Idaho Mining Association. Idaho's second most valuable industrial mineral commodity is construction sand and gravel, worth $46.3 million in 1994 (**Tables 2** and **3**).

*Table 2. Value (in thousands of dollars) of industrial mineral production in Idaho, 1991 to 1994. Values include some estimates of 1994 production. Phosphate value is that of the phosphoric acid and elemental phosphorus products of the mined phosphate ore. See **Table 1** caption.*

Year	Total Industrial Mineral Value	Processed Phosphate Value	Sand and Gravel (construction) Value
1991	$ 664,012	$ 585,695	$ 31,300
1992	639,805	546,842	40,728
1993	676,234	568,152	44,900
1994	$ 728,998	$ 630,000	$ 46,300

Table 3. Value of industrial mineral raw ores produced in Idaho in 1994, expressed as a percent of their total value. Total value of raw ore production in 1994 from Idaho's industrial mineral mines was approximately $253.9 million.

Raw Ore	Percent of Total	Value
Phosphate	60.7	$154 million
Sand and Gravel and Crushed Stone	26.2	
Cement and Lime/Limestone	10.9	
Other (including Garnet, Feldspar, Pumice, Silica, Gems, Clay, Zeolites)	2.1	
Total	**99.9**	**$253.9 million**

According to the Idaho Department of Employment, there are some 6300 mining-related jobs in Idaho, which has a total population of slightly over one million people. Some 1300 persons are directly employed by industrial mineral mines and quarries. Another 2000 persons work in southeastern Idaho's phosphate processing plants, and the remainder work in stone, glass, and clay industries. With an average wage in 1994 of $34,515, mining constitutes one of the state's highest paying occupations.

Geologic setting

Idaho's varied, complex geology has created a diverse suite of industrial minerals. All periods of geologic time are represented in the state. A small piece of Archean age craton is preserved in south-central Idaho, but the largest exposures of Precambrian rocks are in northern Idaho, where the clastic sedimentary rocks of the Proterozoic Belt Basin host the silver-lead-zinc-copper veins of the famous Coeur d'Alene Mining District (**Figure 1**). A wide belt of metamorphosed Precambrian to Paleozoic rocks, including garnetiferous and kyanite-bearing schists border the Cretaceous Idaho Batholith in north-central Idaho. Folded and thrust-faulted Paleozoic and Mesozoic sedimentary rocks form the mountain ranges in east-central and southeastern Idaho, where they host important limestone and phosphate reserves. Granodiorite and related rocks of the Idaho Batholith were emplaced some 70 to 100 million years ago, following accretion of the Seven Devils terrane onto the western edge of the continent. Eocene-age Challis volcanics, associated with several precious metal deposits, cover eroded plutons and sedimentary rocks in east-central Idaho. Volcanism again in the Miocene resulted in outpourings of the Columbia River flood basalts in western Idaho, Oregon, and Washington. These basalts are often crushed and utilized as road material and landscaping rock. Continued uplift and erosion of the batholith resulted in deposition of the late Tertiary silica-feldspar sands of the Boise foothills and many alluvial gravel deposits. Lacustrine sediments and volcanic ash deposits in the ancestral Snake River Plain graben formed beds of zeolites, clay, limestone, and diatomite. Scoria, pumice, and perlite are associated with the progression of Tertiary to Quaternary Snake River Plain volcanism from west to northeast as the North American plate moved over the Yellowstone hotspot. Other Idaho mineral deposits, such as silicified building stone, opal, and jasper are associated with hot spring activity during these periods of volcanism.

This full geologic spectrum of rock types, numerous orogenic and volcanic periods, active seismicity, and hot springs has provided Idaho with a wealth of varied mineral deposits, as well as the state's natural beauty. Numerous publications and maps of the Idaho Geological Survey and other references are available for more detail on Idaho's geologic past.

Phosphate district

The Idaho Phosphate Reserve constitutes the state's most significant mineral deposit, of any type. The phosphate industry is an economic mainstay of southeastern Idaho. Idaho is the third largest phos-

phate producer in the United States and currently the only producer of elemental phosphorus. Four large open pit mines and five processing facilities are in operation near the cities of Pocatello and Soda Springs (**Figures 1** and **2**). FMC Corporation has a mine at Dry Valley and an elemental plant (the largest such facility in the world) in Pocatello; Monsanto Company operates the Enoch Valley mine and an elemental plant at Soda Springs. Rhone-Poulenc Basic Chemicals Company extracts ore from their Rasmussen Ridge mine; the ore goes to Agrium's fertilizer plant at Conda, Idaho. The J.R. Simplot Company mines phosphate from the Smoky Canyon property to feed their fertilizer plant at Pocatello. The mines are all located in Caribou County near the Idaho-Wyoming border (**Figure 2**). In 1995, the four open pit mines produced an estimated 6.75 million tons of phosphate ore, having a raw ore value of $125 million, but a processed ore value of approximately $664 million. Processing plants have been aided by Idaho's historically cheap hydropower and good transportation and infrastructure in the area.

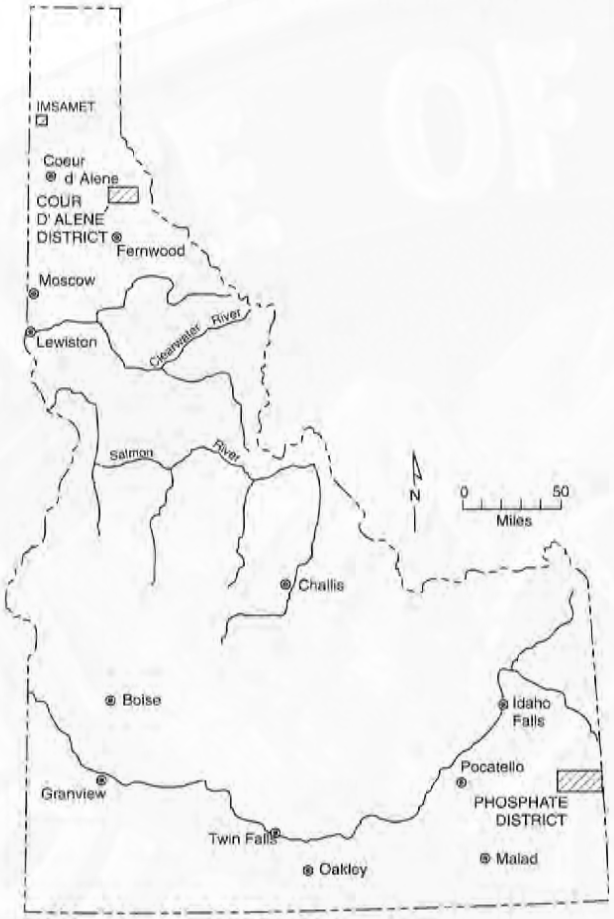


Figure 1. Location of selected cities and mining districts in Idaho.

The phosphate ore is hosted by the Meade Peak Member of the Pennsylvanian-Permian Phosphoria Formation. The original Phosphoria basin extended over what are now parts of Idaho, Utah, Wyoming, and Montana. Southeastern Idaho lies within the Overthrust Belt, and the Phosphoria Formation beds have been affected by Cretaceous-age folding and thrust faulting, which locally repeats and offsets the units. Most of the ore horizons currently being mined are tilted at a steep angle. Ore consists of brown to black phosphatic cherts and shales. Apatite is the principal ore mineral. Two ore seams are mined, with a middle waste unit located between them. Ore reserves are typically in zones where weathering or supergene effects have leached and altered some of the rock constituents, resulting in higher grade ore averaging 20 to 30% P_2O_5.

All the ore mined in Idaho is now processed in-state into two basic products: phosphoric acid for fertilizer production or elemental phosphorus. To produce phosphoric acid, the ore is washed, ground, and calcined (at about 1450°F) to remove water and organic matter. Calcined ore is reacted with sulfuric acid to produce phosphoric acid (H_3PO_4) of several grades and concentrations. A principal product is green, liquid superphosphoric acid (about 70% pure). Dry granular fertilizers are also made after reacting the phosphoric acid with ammonia. Gypsum is a manufactured waste product of the process. Fertilizers are shipped principally to the Midwest and Western United States.

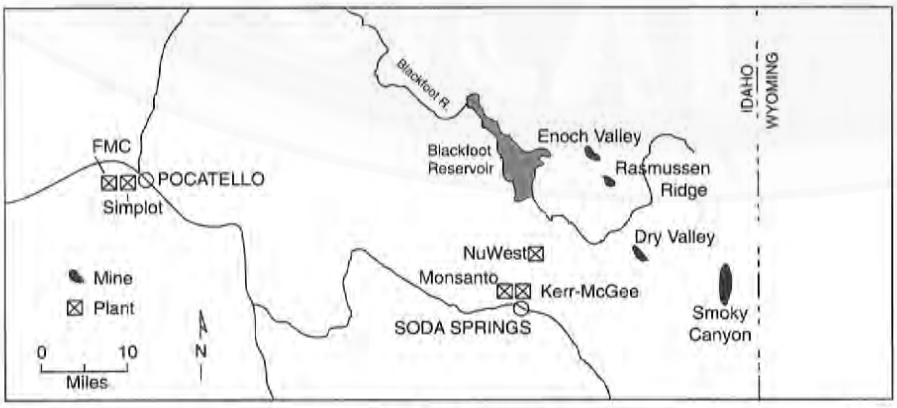


Figure 2. Location of mines and plants in southeastern Idaho's phosphate district, 1995.

Elemental phosphorus production is a different process. Crushed ore is calcined and mixed with coke and silica (from quartzite quarries), then melted in an electric furnace at temperatures over 2500°F. The phosphorus and other el-

ements are volatilized, and the gas is cleaned and condensed into liquid elemental phosphorus, P_4. The heavy, molten ferrophosphorous slag is cooled and can be sold for its vanadium and other metal content.

FMC opened the Dry Valley mine in 1992 (**Figure 2**), following closure of the Gay mine. In 1995, the original A pit was being backfilled and reclaimed as production moved south into the B pit. Approximately 1.3 million tons of phosphatic shale were mined in 1995, and drilling was in progress to continue mining the B pit southward as planned. Mining is done by a shovel and large backhoe, in combination with two front-end loaders. The Dry Valley mine has 80 employees and hosts a particularly innovative, computer-automated ore loading system (**Figure 3**). Stockpiled ore is blended and pushed into an underground crusher, goes up a conveyor, through an automatic sampler and into an automatic hopper. The computerized hopper drops a certain weight of ore into a level grade railroad car, set into position by a computer-controlled car puller. The lack of slope on the railroad tracks and the computerized car movements greatly decrease the chance of human injury and allow only four persons to monitor the entire operation. The computer also keeps the shipping records.



Figure 3. Automated railroad-car loading hopper at FMC's Dry Valley phosphate mine, Caribou County, Idaho.

FMC's ore train goes to their elemental phosphorus plant in Pocatello. The plant employs over 500 persons and is the largest such plant in North America with a capacity of 270 million pounds of phosphorus annually. Numerous technological and environmental improvements have been implemented over the past few years, including the addition of the world's largest stacker/reclaimer unit (a large, yellow wheel that blends the stockpiled ore). FMC uses and sells the elemental phosphorus for a variety of products, including soft drinks, baking powder, cereals, toothpaste, cleaning products, light bulbs, tires, matches, and many other everyday items.

Monsanto operates the Enoch Valley mine (**Figure 2**), which produces 1.2 million tons of ore a year from a combination of state and Federal leases. Mining contractor, Dravo, uses Hitachi shovels and Caterpillar 85-ton trucks in the mining (**Figure 4**). At the tipple, 70 tons of ore is loaded into each trailer of a triple trailer truck. The triple trailer truck, which has a gross vehicle weight of 315 tons, then transports the ore on a 19-mile-long, private runway-like road to the plant near Soda Springs. Monsanto also operates a quartzite quarry to supply silica feed to the plant.

Monsanto has been a leader in the reclamation business, pioneering ways of promoting wildlife habitat in the reclaimed pits, reseeding and planting seedlings on the recontoured dumps, and starting growth in constructed wetlands. Among many awards, the company received a 1994 Idaho Land Board award for environmental excellence, as did the company's environmental coordinator at the mine. Monsanto also received an award from the Wildlife Habitat Council of Maryland for its innovative environmental programs.

The company had an active exploration program in 1995 at the South Rasmussen property. All of the phosphate companies are awaiting the results of an Environmental Impact Statement (EIS) being written for the Caribou National Forest on two proposed lease areas, Manning Creek and the Dairy Syncline. After the EIS is completed, the areas may be put up for competitive bid.

Ore from Enoch Valley is sent to Monsanto's elemental phosphorus plant in Soda Springs. Annual production is over 200 million pounds of phosphorus. Much of that is manufactured on-site into Monsanto's popular weed killer, Roundup. One of the waste products of the elemental process is ferrophosphorous slag. Monsanto sells the slag to the nearby Kerr-McGee vanadium plant. Vanadium, as well as uranium and other elements, is naturally enriched in the phosphatic ores.

The Kerr-McGee vanadium plant has operated for 15 years without a lost time accident, a record the company and employees are very proud of. Forty tons of ferrophosphorous slag and an equal amount of limestone are used daily as raw material for the vanadium plant. The slag mixture is roasted at 1500°F and goes through various chemical processes to end up as fused, pure vanadium oxide or special products like SAVAN, an orange-colored, sodium ammonium vanadate powder. The vanadium is used in steel alloys and catalysts.

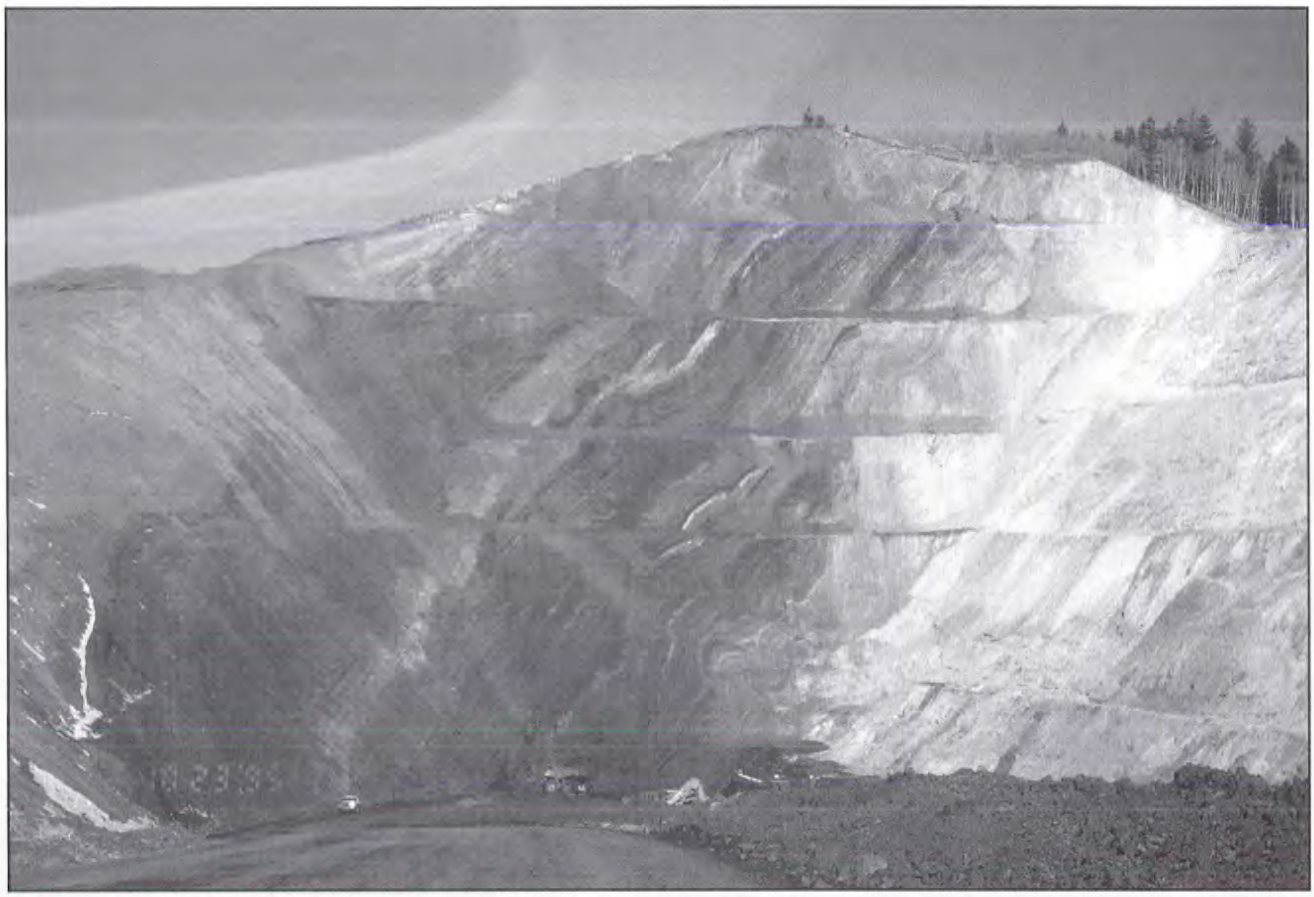


Figure 4. Folded and faulted Phosphoria Formation exposed in the South Pit of Monsanto's Enoch Valley phosphate mine, Caribou County, Idaho.

J.R. Simplot Company, a Boise-based private firm best known for its food processing operations, mines phosphate ore at the Smoky Canyon mine (**Figure 2**), located near the Wyoming border. Using 25-cubic-yard shovels and 170-ton Haulpacs, the mine extracts 2.5 million tons of ore annually, making it the state's largest phosphate producer. In 1995, engineers were designing the Panel D pit, and a new group of exploration geologists were drilling on the Freeman Ridge exploration license. Crushed ore is pumped through an 87-mile-long slurry line to a fertilizer plant in Pocatello (**Figure 5**). Simplot employs some 500 persons at the plant, which produces approximately 400,000 tons of fertilizer annually.

Rhône-Poulenc Basic Chemicals Company acquired the holdings of Stauffer Chemical's phosphate operations in 1987. After the Wooley Valley mine was mined out, the Rasmussen Ridge mine (**Figure 2**) opened in 1991, shipping 600,000 tons of ore annually to Rhône-Poulenc's elemental phosphorus plant in Silver Bow, Montana. The Silver Bow plant closed on October 1, 1995, and all Idaho phosphate ore is now processed within the state boundaries. In 1994, Rhône-Poulenc and NuWest Industries negotiated a 7-year contract to expand production at Rasmussen and supply NuWest's acid plant at Conda. Consequently, all 1.7 million tons of Rhône-Poulenc's ore is now trucked several miles and then loaded into railroad cars for the trip to Conda, a few miles north of Soda Springs. The Rasmussen Ridge mine has 110 employees and like the other mines, it has an active concurrent reclamation program. The open pit covers about 2500 feet of the over 20,000 feet of strike length of the Phosphoria Formation on the lease.



Figure 5. Simplot's phosphoric acid plant at Pocatello, Idaho.

NuWest Industries operates a phosphoric acid fertilizer plant at Conda, having bought out its partners (the Conda Partnership) in 1991. NuWest was sold in the fall of 1995 to Agrium, Inc., a growing Canadian agricultural chemical firm with potash mines and ammonia plants in Canada and the United States. The NuWest plant employs 300 people and receives its ore feed from Rhone-Poulenc's Rasmussen Ridge mine. The plant produces green superphosphoric acid. It recently closed one of its two sulfuric acid plants and was building four storage tanks to receive sulfuric acid purchased from Kennecott's Utah copper smelter near Salt Lake City.

Like others in the mineral industry, Idaho phosphate companies have greatly improved their environmental practices. Past problems at the processing plants have resulted from calciner emissions, water quality, and the slight radioactivity of by-product slag, which was formerly used as aggregate for buildings and roads. Monsanto, FMC, and Simplot have spent millions of dollars complying with new, more stringent air-quality standards. Elemental phosphorus production is particularly energy intensive, and rising energy costs, as well as the environmental costs, are disconcerting to the industry. Mine-site reclamation efforts have been particularly rewarding. Pits are routinely reclaimed, and the landforms resloped and reseeded concurrent with mining elsewhere on the property. Wildlife and shrub habitat are restored and enhanced. The final result is that the closed minesites appear almost undisturbed.

Other industrial minerals

Figure 6 shows the location of Idaho's many other industrial mineral operations active in 1995. While smaller than the phosphate mines, they are widely distributed throughout rural Idaho. Many have been in business for a long time and are a stable and important segment of the rural economies. A mineral-related plant is the IMSAMET aluminum can recycling plant at Hauser Lake in northern Idaho (**Figure 1**).

Garnet

Idaho's largest placer operation is near Fernwood in northern Idaho (**Figure 1**). In 1995, Emerald Creek Garnet (**Figure 6**) produced about 27,500 tons of industrial placer garnets from stream gravels and weathered garnet-bearing schists. Idaho is the largest garnet producer in the nation because of the Emerald Creek operations. The local company was bought by a Canadian firm, Western Garnet Company, in 1992. The pink almandine garnets are dredged from Emerald and Carpenter Creeks (**Figure 7**). Last year the firm made improvements to its two concentrating mills and expanded their reclamation program. The garnets, available in over 30 different sizes, are used principally in abrasives, filters, and water jet cutters.

Limestone and cement

Current limestone operations are located in southern Idaho (**Figure 6**). Rock is mined from Paleozoic carbonates in the southeast or Tertiary deposits in the southwestern desert. Idaho Lime did build a filler-lime plant in Grangeville in northern Idaho several years ago, but the facility is currently idle. The Mission Creek deposit, operated by the Nez Perce Indian tribe, and other deposits have had only minor activity the past several years. Faxe Kalk, a Danish firm, made headlines in 1995, when Secretary of Interior Babbitt pro-

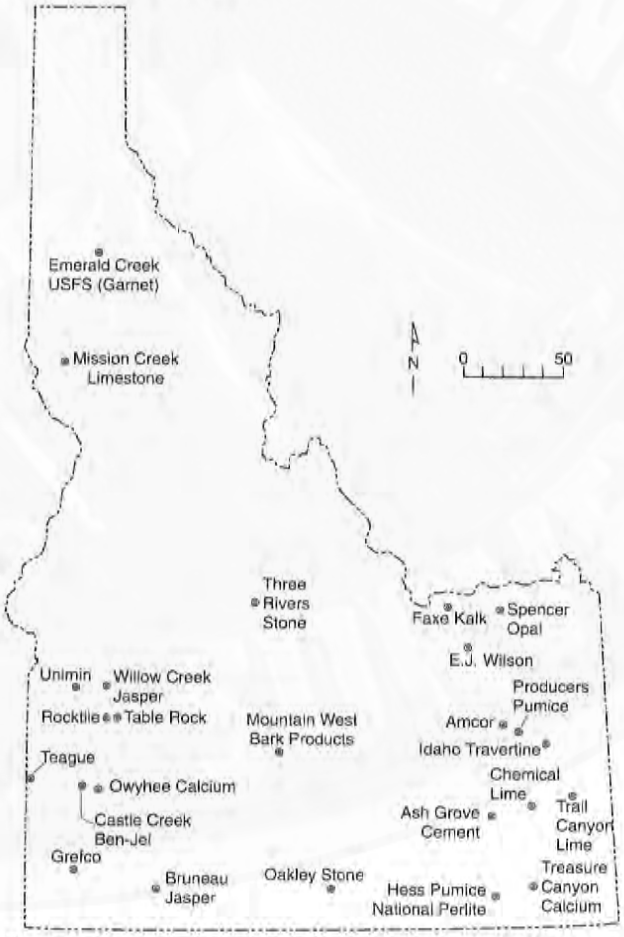


Figure 6. Location of industrial mineral operations (except phosphate) in Idaho, 1995.



Figure 7. Emerald Creek Garnet's mine, Fernwood, Idaho.



Figure 8. Maerz kiln at Chemical Lime plant, Ten Mile Pass, Caribou County, Idaho.

tested the signing of patents on 108 acres of travertine at Lidy Hot Springs west of Dubois in Clark County. Faxe Kalk bought the claims from a local firm, E.J. Wilson and Sons, who still quarry dolomite and limestone for cattle feed supplements, important to southern Idaho's expanding dairy industry. Faxe Kalk has been conducting an aggressive marketing and testing program on the high grade travertine, which may be used as a filler for the paper industry.

Ash Grove Cement in Inkom, just south of Pocatello, is Idaho's only cement producer. In 1995 they had an excellent year, shipping 230,000 tons of cement. Limestone is quarried from near the plant site. Recent exploration drilling and seismic work showed that the limestone continues at depth, providing a large reserve. The company installed a used tire burner in 1993 on one of its kilns and is thinking of adding a second unit.

The Chemical Lime (formerly Chemstar Lime) plant and quarry near Bancroft in Caribou County is the state's newest limestone operation. It opened in 1992. Crushed limestone from the quarry is carried on a half-mile long conveyor to an energy efficient, 171-foot-high Maerz kiln, which heats the rock up to 220°C, driving off the CO_2 and leaving calcium oxide (**Figure 8**). The facility operates 24 hours a day with 40 employees. The lime is used for pH control in gold mines, in steel mills, water purifiers, and other uses.

Two companies, Castle Creek Mines and Owyhee Calcium Products, mine oolitic limestone from Tertiary lake bed deposits in Owyhee County of southwestern Idaho (**Figure 6**). Castle Creek mines about 6000 tons a year and Owyhee Calcium a somewhat larger tonnage from pits south of Grandview. The material has been used for cattle feed supplements, but a newer use is for acid-control in gold mines and emergency acid spill clean-up kits.

Treasure Canyon Calcium, in southeastern Idaho shipped about 20,000 tons of limestone from quarries near Preston (**Figure 6**) to Simplot's acid plant at Pocatello. The company employs 5 people. Also in southeastern Idaho, Vaughn Smith Construction mined and shipped about 10,000 tons of limestone from the Trail Canyon mine north of Soda Springs to Kerr-McGee's plant. Kerr-McGee uses the stone for processing vanadium products and is Vaughn Smith's sole customer for lime.

Clay

The same sequences of Tertiary lacustrine deposits which contain the oolitic limestones also host bentonite clay beds and zeolites. Unimin Corporation took over the Castle Creek bentonite mine from AIMCOR in 1994. Other Idaho clay producers include Mutual Materials in northern Idaho and E.J. Wilson and Sons in eastern Idaho.

Zeolites

Zeolites occur in the Castle Creek area and also a few miles west near the Oregon border, where Teague Mineral Products mines several thousand tons of potassic clinoptilolite from altered Tertiary ash beds. The company recently added a new jaw crusher to their plant in Adrian, Oregon. Zeolite is used in cation exchange processes, including radioactive element absorption, wastewater treatment, and odor control.

Diatomite

An inactive, but potentially important resource is Grefco's Deep Creek diatomite deposit in southernmost Owhyee County. The company has been evalu-

ating it for several years. In 1992, the Bureau of Land Management conducted a drilling and sampling program during a patent examination. Part of the procedure included lowering a geologist a short distance down a 30-inch diameter drill hole to sample the diatomite layer. Another large diatomite resource is the Clover Creek deposit in Gooding County.

Silica and feldspar

Unimin Corporation has been producing industrial grade silica sand for abrasives and glass-making from pits near Emmett, 20 miles northwest of Boise, for many years. One of their recent specialty products is sand for golf course bunkers. According to a recent press article, Idaho's pure white sand is being sold to golf courses in Hawaii and Canada. Connecticut-based Unimin is cited as a leading supplier of bunker sand in the country, but only 2 of its 40 plants can supply the high-quality product. The Emmett deposit is a Tertiary lake-bed sand, composed of quartz and feldspar eroded from the Idaho batholith. Unimin reclaims the feldspar as another industrial mineral product from its Emmett plant.

Volcanic rock deposits

Most of Idaho's pumice, scoria, and lava rock production comes from the abundant Tertiary to Recent basaltic to rhyolitic volcanic rocks of southern Idaho (**Figure 6**). Mountain West Bark, a division of Hecla Mining Company, operates several pumice and scoria pits at Fairfield, Rexburg, and other localities. Three colors are produced for landscaping: gold rock, black rock, and red rock. About 60,000 cubic yards were mined in 1995, with the red lava rock being the most popular. It is quite striking on landscaped lawns and gardens.

Hess Pumice mines pumice at the Wrights Creek mine in southeastern Idaho and processes it at a plant in Malad (**Figure 9**). While some of the material is used for lightweight aggregate, Hess also makes a specialty product, an ultrapure, ultrafine, low heavy mineral pumice powder used for grinding television screens and other applications. The specialty pumice is exported to the Pacific Rim countries and Europe. Demand for this product has steadily increased. In 1995 the company shipped 90,000 tons of pumice, and was building a second, fine-grinding facility. Other pumice producers in the state include Amcor in Idaho Falls and Producers Pumice with a mine in Bonneville County and plant in Meridian. Most of this rock is used for lightweight aggregate.

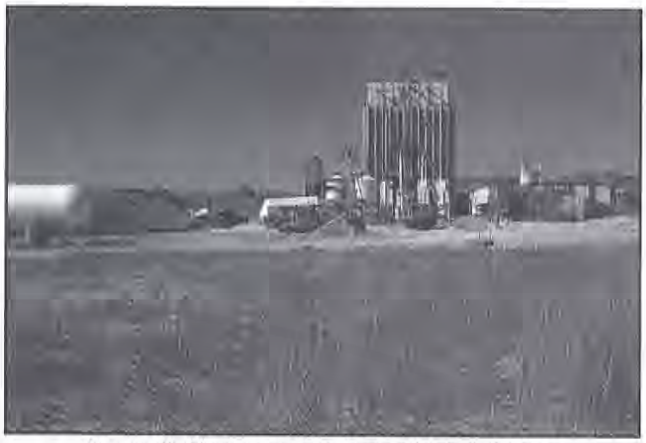


Figure 9. Hess pumice plant at Malad, Idaho.

National Perlite, a subsidiary of Oglebay Norton, operated a perlite mine and processing plant in Malad until 1992, when it was shut down due to air quality problems. The company has been up for sale. Perlite is used in fireproofing, filters, and lightweight aggregate.

Decorative stone

As Idaho and the neighboring states have prospered in recent years, market demand for decorative stone, whether for dimension stone or landscaping rock, has increased. Three Rivers Stone quarries an argillaceous quartzite near Clayton (**Figure 6**). Increasing demand allowed the company to add new machinery at the site. The attractive stone went to fancy houses in Ketchum and Kirk Douglas' home in California.

Oakley stone, a unique micaceous quartzite, is Idaho's most famous dimension stone. Available in tan, gold, and gray colors, it splits easily into thin sheets which can be used for a paving or facing stone. It is quarried south of the town of Oakley (**Figures 1** and **6**) by three Idaho operators, Oakley Valley Stone Inc., Northern Stone Supply Inc., and Dave Rodriguez. Demand is good, but international competition is increasing.

A Boise firm, Rocktile Specialty Products, Inc., has operated a stone cutting facility for 15 years, producing stone tiles in a variety of colors, sizes, and types. Oakley stone, from Idaho and adjacent Utah, comprises much of their raw material. The company has been increasing their export market.

Table Rock sandstone, quarried from Table Rock Mesa above Boise, is another historic Idaho building stone. Silicification along Tertiary(?) age faults bounding the western Snake River Plain cemented loose sands into a dense, hard sandstone which forms

prominent ridges and mesas along the capital city's skyline. In the early 1900s, prisoners quarried the sandstone for building the Old Penitentiary and providing stone for the Idaho Statehouse, Boise churches, and buildings as far away as San Francisco, California, and Vancouver, British Columbia. Today, Table Rock Sandstone, Inc. quarries new material to meet special orders for buildings and landscaping rock. For example, in 1993, columns of sandstone were installed as part of the new Fine Arts Building on the campus of Albertson College of Idaho (**Figure 10**). The columns were turned by Idaho Travertine. Approximately 1200 tons of sandstone went into the building.

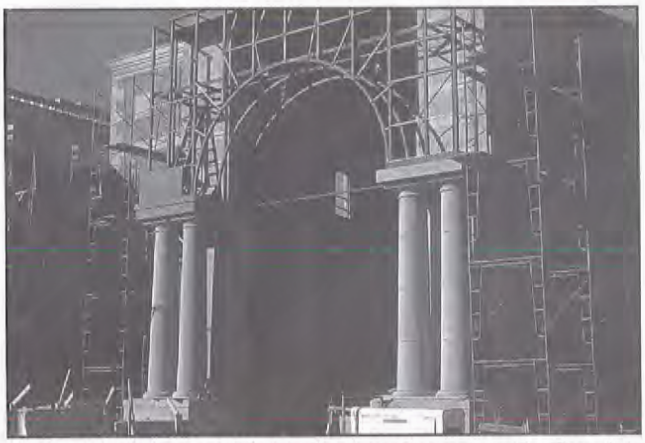


Figure 10. Columns of Table Rock sandstone on the Fine Arts Building at Albertson College of Idaho, Caldwell, Idaho.

Idaho Travertine in Idaho Falls is one of the largest stone fabricators in the West. They quarry travertine at the Fall Creek mine, but cut stone from across the country with their oversized diamond gang saws. Recent projects have included the new Idaho Power building in Boise, and cutting the sandstone columns for the Fine Arts Building at Albertson College of Idaho in Caldwell.

Population growth and a robust economy in the Boise River Valley and other populated centers in Idaho, such as Coeur d'Alene, Sun Valley, Pocatello, and Twin Falls, has fueled a booming business in landscaping rock for residential as well as commercial establishments. Lichen-covered sandstone was particularly prized for a while, and slabby rhyolite, river rock, sandstone, and polished basalt have all been in fashion lately. The Bureau of Land Management is having increasing problems with mineral trespass caused by the removal of landscaping rock and decorative stones from unpermitted locations on Federal land.

Sand and gravel

Idaho's production and need for sand and gravel has accelerated over the last few years due to growth in most urban areas, such as Boise, Ketchum, and Coeur d'Alene. In 1994, Idaho produced 15.95 million tons of construction sand and gravel, up 6.6% from the previous year, according to the U.S. Bureau of Mines. In the Boise area, and elsewhere in the state, construction demand is high. Gravel pits are located in recent alluvial gravels of the Boise River and several old stream terraces as much as 1 million years old (Othberg, 1994; Othberg and Stanford, 1992). Operations making concrete (**Figure 11**) produce clean gravel from the active floodplain, but operators are finding that the older terrace gravels (**Figure 12**) are not always suitable for the same uses. Existing supplies and potential resource areas are becoming surrounded by residential developments, whose occupants object to the noise and dust generated by these operations.



Figure 11. Monroc Concrete's plant at Eagle, Idaho.



Figure 12. Gowen Terrace Gravel overlain by Fivemile Basalt (age about 1 million years) near Boise, Idaho.

Gemstones

Idaho is nicknamed the "Gem State," as the word Idaho was supposedly derived from an Indian word meaning "Gem of the Mountains." Nevertheless, the nickname is an appropriate reference to the state's many mineral treasures. Rockhounding and gold panning are popular hobbies. Some Idaho gems are also produced commercially. Silicified Tertiary rhyolite of variegated colors and jasper are mined from the Willow Creek jasper mine, located north of Eagle, and the Bruneau jasper mine south of Bruneau in southwestern Idaho. Aquamarines, topaz, and smoky quartz crystals are known from several Tertiary-age batholiths, though collecting in some areas like the Sawtooth Wilderness is prohibited. One company, Golconda Resources, has been exploring for diamonds in the McCall area, where three diamonds were reportedly found in a placer in the 1930s. Precious opal is mined near the town of Spencer in Clark County of eastern Idaho and is widely used in jewelry. The opal occurs as thin layers within partially filled gas cavities in a rhyolite flow.

The Star Garnet is the state gemstone. Found only in Idaho and India, star garnets display four or six rays (**Figure 13**). The purplish-red, gem garnets and excellent mineral specimens are found in northern Idaho in Latah and Benewah Counties, near the Emerald Creek industrial garnet operation. During the summer, the U.S. Forest Service operates a garnet dig near St. Maries, where tourists can pay a modest fee and dig for their own gems. Some do take home a star garnet treasure, the rest just fond memories of a day spent exploring for minerals in Idaho.



Figure 13. Idaho Star Garnet and euhedral garnet in schist. Courtesy of Stewart's Gem Shop, Boise, Idaho. Photograph is slightly smaller than actual size (jewelry box is about 4" long by 3" wide).

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Industrial rock and mineral resources and developments in Utah

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Abstract

Utah produces a wide variety of industrial rock and mineral commodities, valued in 1995 at $429 million (17% of the total non-fuel mineral production of Utah). Commodities produced in large amounts include: halite; non-halite salts; Portland cement; carbonates including lime, limestone, and dolomite; sand and gravel; phosphate; crushed stone; clays; and gypsum. Commodities produced in smaller amounts include: building stone, gemstones, silica, perlite, and scoria.

The past three years have been a period of slow, steady growth in industrial rock and mineral production. However, some commodities have grown much more rapidly than others. The robust economy of the West has stimulated production of construction-related commodities like sand and gravel, Portland cement raw materials, lime, and building stone, resulting in exploration for raw materials, development of new quarries, and expansion of existing plants. Noteworthy specific developments include: 1) strong exploration efforts for sand and gravel, limestone, high-alumina clay, and red beryl; 2) the opening of three new limestone quarries; and 3) expansion of a lime operation and planned expansion of two cement plants.

Introduction

Although the economy of Utah has diversified in recent years, mineral production, including industrial rocks and minerals, remains an important segment of the state's economy. Minerals-industry products were valued at $2.5 billion in 1995; $429 million of this amount came from industrial rocks and minerals production (Bon and others, 1996). The industrial mineral commodities with the highest production values in Utah in 1995 were: halite, potash, and other salts; Portland cement; lime, limestone, and dolomite; sand and gravel; phosphate; crushed stone; clays; and gypsum. Locations of the companies producing these commodities are shown in **Figure 1**.

The purpose of this report is to summarize: the industrial mineral industry in Utah, which companies produce what commodities from which geologic formations, and what changes are taking place in the industry. Minor, intermittent mineral production by very small companies is not addressed. Utah humate and gilsonite production is also not included in this report since they are not classified as industrial minerals but are lumped genetically with Utah fuel resources. This report supersedes an earlier published report on Utah industrial minerals (Tripp, 1991).

Halite

Significant halite resources occur at seven locations in Utah (**Figure 1**): 1) in surface brines of Great Salt Lake, 2) in bedded salts and brines of the Pennsylvanian Paradox Formation of the Hermosa Group of the

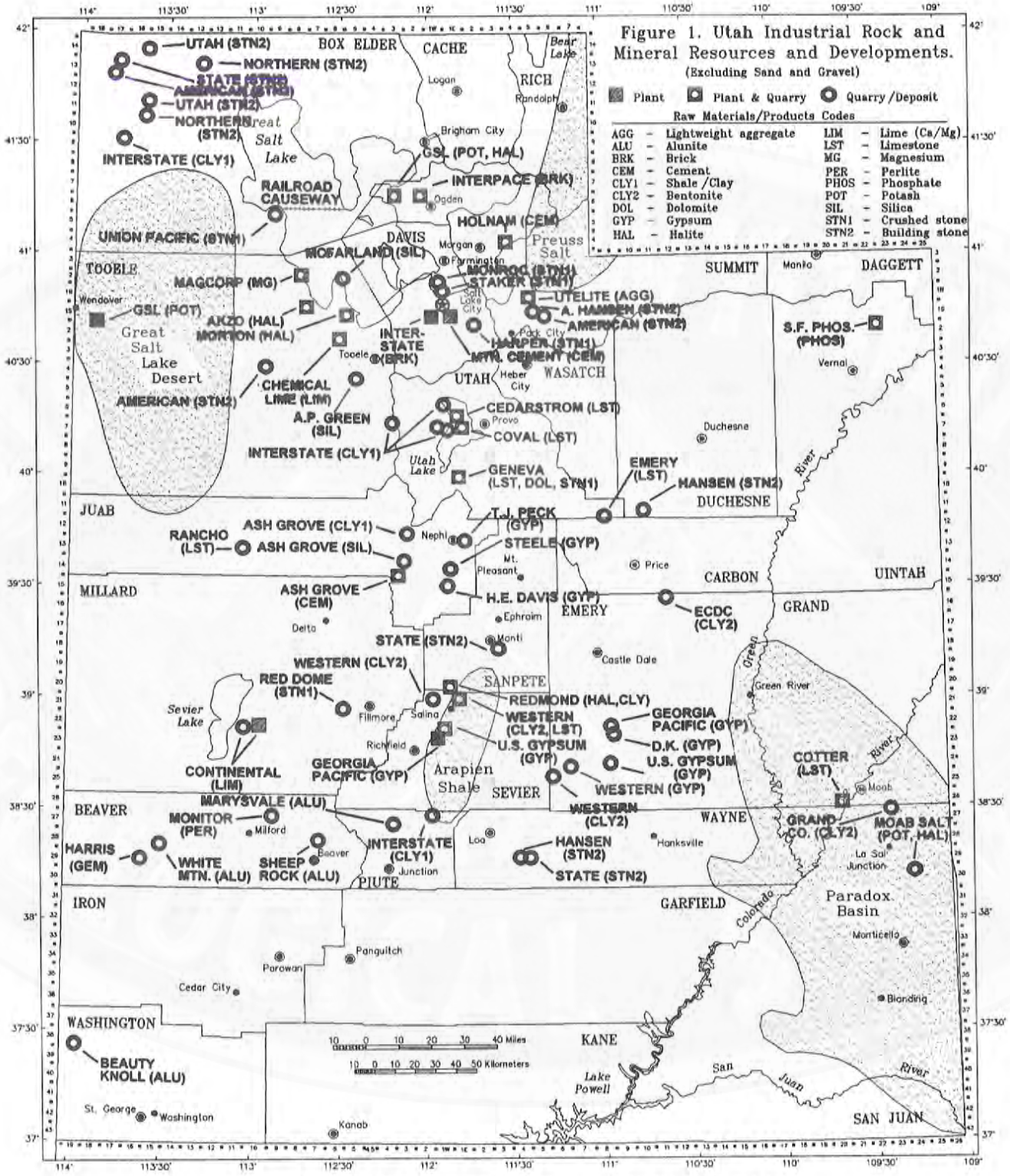


Figure 1. Industrial rocks, mineral resources, and developments in Utah.

Paradox Basin, 3) in subsurface brines in Holocene and Quaternary sediments of the Great Salt Lake desert, 4) in bedded salts of the Jurassic Arapien Shale of Sevier and Sanpete Counties, 5) in subsurface brines in Holocene sediments of Sevier Lake, 6) in Tertiary salt domes of Millard County, and 7) in bedded salts of the Jurassic Preuss Sandstone of northeastern Utah.

Great Salt Lake

Great Salt Lake has long been an important producer of halite by solar evaporation of surface brines in shallow harvest ponds. Great Salt Lake production accounted for the bulk of the 2.6 million tons [2.4 million metric tons (mt)] of salt produced in Utah in 1995 (Bon and others, in press). The lake consists of two distinct brine bodies, the northern and the southern parts of the lake, which are separated by the Southern Pacific railroad causeway. In July 1996, the southern part of the lake had a brine density of 1.086 g/cm^3 (grams per cubic centimeter) and is becoming less salty since it receives fresh water from the two main rivers entering Great Salt Lake. The northern part of the lake had a July 1996 brine density of 1.22 g/cm^3 and is saturated. Halite is precipitating in the northern area of the lake (Gwynn, 1995; Gwynn, 1996, verbal communication). The lake brine contains commercial concentrations of sodium, potassium, and magnesium salts but is not anomalously rich in other salable commodities such as lithium, bromine, and boron (Sturm, 1980). Three companies currently produce halite from Great Salt Lake: Great Salt Lake Minerals and Chemicals Company/North American (GSL); Morton Salt Company; and Akzo Salt, Inc. North American and Great Salt Lake Minerals and Chemicals Company are divisions of Harris Chemical Group, Inc. and share some production facilities.

Other halite deposits

The other halite deposits in Utah contain large resources but have produced small amounts of halite compared to Great Salt Lake. In the Paradox Basin near Moab, Moab Salt, Inc. (acquired in 1995 by Potash Corporation of Saskatchewan) recovers halite as a by-product of potash mining. They solution mine bedded evaporite salt of the Pennsylvanian Paradox Formation (salt cycle five) and process the resulting brine by solar evaporation. Reilly Industries' potash plant in the Great Salt Lake desert near Wendover has generated large amounts of halite as a by-product of potash production. Most of the harvested halite still fills the evaporation ponds where it precipitated during a preliminary step in potash recovery; only small quantities of the halite have been shipped to market. In central Utah, the Redmond Clay and Salt Company mines rock salt underground from a diapir in the Jurassic Arapien Shale. This rock salt is primarily sold for livestock salt and marketed as a health-food table salt. Crystal Peak Minerals attempted to produce halite from subsurface brines of the Sevier Lake playa but the operation was discontinued in 1993 due to lack of a market for the salt. Near Delta, oil and gas drilling at the Argonaut Energy No. 1 Federal well revealed the presence of a salt dome in Tertiary rocks. The salt section in this well is more than 5000 feet (1524 m) thick (Mitchell, 1979). This deposit has not been developed. In northeastern Utah, the Jurassic Preuss Sandstone contains a poorly known and presently undeveloped salt resource.

Non-halite salts

Four localities in Utah contain large quantities of potassium and magnesium salts and sodium sulfate: Great Salt Lake, the Great Salt Lake desert, the Paradox Basin, and Sevier Lake. In addition, there are large alunite deposits in southwestern Utah that are potential sources of both potash and aluminum.

Great Salt Lake

The two Great Salt Lake operations that produce salts other than halite are: the Magnesium Corporation of America (Magcorp), which produces magnesium chloride brine for conversion to magnesium metal and chlorine gas, and GSL, which produces potassium sulfate (potash) and magnesium chloride. The potash resource of Great Salt Lake is estimated to be 100 million tons (91 million mt) of K_2O equivalent (Adams and Hite, 1983). An additional saline resource, buried at shallow depth within the Quaternary sediments near the east end of the Great Salt Lake railroad causeway, consists of a bed of mirabilite (a hydrated sodium sulfate) which reaches a maximum thickness of about 32 feet (10 m). Construction crews discovered the mirabilite while building the original wooden railroad trestle across Great Salt Lake in 1903 (Hite, 1964).

Great Salt Lake desert

Reilly Industries' Wendover potash plant produces potassium chloride (potash) and magnesium chloride brine from brines of three subsurface aquifers at the Bonneville Salt Flats in western Tooele County. The shallowest aquifer, generally less than 20 feet (6 m) deep, provides most of the brine which is gravity drained through canals toward the plant. Brine "elevators" raise the brine from the canals to the solar evaporation ponds. A froth flotation circuit separates the valuable sylvite from the sylvinite (a sylvite/halite salt mixture) harvested from the solar ponds. Less concentrated brine from a deep aquifer is produced from wells as deep as 2051 feet (625 m). Wells as deep as 200 feet (61 m), in an alluvial-fan aquifer to the north of the Wendover plant, provide brackish water used in the plant for processing sylvite (Bingham, 1980). The potash resource of the Great Salt Lake desert is estimated to be 10 million tons (9 million mt) of K_2O equivalent (Adams and Hite, 1983).

Paradox Basin

Bedded sylvite, carnallite, and associated subsurface brines underlie a large part of the Paradox Basin. The potash resource occurs within 18 of 29 evaporite cycles in the Pennsylvanian Paradox Formation of the Hermosa Group; eleven of these cycles contain significant amounts of potash (Hite, 1961). The potash resource of the Paradox Basin is estimated to be 280 million tons (254 million mt) K_2O equivalent (Adams and Hite, 1983). Moab Salt, Inc. produces about 118,000 tons per year (TPY) [107,000 metric tons per year (MTPY)] of potash by solution mining the old workings of their formerly underground mine (Phillips, 1975). They produce potash from an 11-foot (3-meter)-thick sylvite bed in the fifth evaporite cycle down from the top of the evaporite sequence. This bed occurs at depths of less than 4000 feet (1220 m) (Hite, 1964). Process water for the operation comes from the Colorado River.

Sevier Lake

Sevier Lake, dry through most of historical time, contains subsurface brines comparable to those of Great Salt Lake (Gwynn, 1986), although Sevier Lake brines have a higher sulfate-to-chloride ratio and a lower magnesium content (Whelan, 1969). Crystal Peak Mineral Corporation's defunct Sevier Lake halite operation had planned to eventually add a potash circuit to their plant.

Alunite deposits

The large alunite deposits of southwestern Utah represent an unconventional potash and aluminum resource. These deposits are the largest of their type in the United States. The White Mountain replacement deposit contains an estimated resource of 232 million tons (210 million mt) of ore containing 33% alunite with an additional 402 million tons (365 million mt) of ore containing 28% alunite (Hall, 1978). A consortium of companies planned to produce aluminum and potash from this deposit in the 1970s but decided that the economics at the time were unfavorable. The Marysvale vein and replacement deposits were the sites of minor potash production in World War I and World War II.

Portland cement

Utah contains vast amounts of the raw materials needed for Portland cement production, including high-calcium limestone, natural cement rock, high-silica quartzite and sandstone, clay and shale, iron ore from near Cedar City or industrial by-product iron, and gypsum from Jurassic rock units of central Utah.

There are two current producers of Portland cement: Holnam, Inc. in Morgan County and Ash Grove Cement Company in Juab County. A third cement plant previously owned by Lone Star Industries, Inc. in downtown Salt Lake City, has been inactive since 1988 and was purchased by Mountain Cement Company for use as a cement shipment terminal for production from their Laramie, Wyoming and Fernley, Nevada plants.

Holnam uses limestone from the Jurassic Twin Creek Limestone, a natural cement rock, at its 350,000-TPY (317,000-MTPY) wet-process plant. Other materials utilized include silica from the Triassic-Jurassic Nugget Sandstone, gypsum from the Jurassic Arapien Shale south of Nephi, and by-product iron from Kennecott Copper Corporation near Salt Lake City. Fuel used at the Morgan County plant is primarily coal with natural gas as a backup. Holnam recently announced plans to rebuild their plant as a 700,000-TPY (635,000-MTPY) dry-process plant.

Ash Grove Cement Company uses limestone from the Cambrian Dome Formation adjacent to their 825,000-TPY (749,000-MTPY) dry-process, coal-

fired plant; shale from the County Canyon quarry; and silica from the Permian Diamond Creek Sandstone at the Nielson quarry. The County Canyon and Nielson quarries are located within a few miles of Ash Grove's plant. Ash Grove obtains iron from Kennecott slag and Nucor mill scale. Gypsum for retarding the setting time of the cement is obtained from the Jurassic Arapien Shale at the T.J. Peck quarry. Ash Grove increased their plant capacity from 650,000 TPY (590,000 MTPY) in the spring of 1996.

Lime, limestone, and dolomite

Cambrian to Mississippian geologic formations provide most of Utah's carbonate production. Calcite veins and oolitic sands of Great Salt Lake are other sources of present or past production. Ten operators in Utah produce a wide variety of products:

1) Continental Lime, Inc., located east of Sevier Lake, produces about 730,000 TPY (662,000 MTPY) of high-calcium quicklime in three rotary kilns. The Cambrian Dome Formation provides the feed for the plant.

2) Chemical Lime of Arizona can produce roughly 90,000 TPY (82,000 MTPY) of dolomitic quicklime and hydrated lime from the Ordovician Fish Haven Dolomite. In 1995 they purchased the old Utah Marblehead plant in Tooele County from U.S. Pollution Control, but are not currently operating it.

3) Geneva Steel of Utah quarries their own dolomite and limestone, from a Cambrian carbonate section, for use as flux at their Orem steel mill. Geneva also ships limestone powder to the coal mines of central Utah for rock dust and crushes stone for aggregate (Hawes, 1992).

4) Cotter Corporation mines roughly 25,000 TPY (23,000 MTPY) of limestone from the Pennsylvanian Hermosa Group in San Juan County for flue-gas desulfurization at the Nucla, Colorado power plant.

5) Emery Industrial Resources began producing limestone for coal-mine rock dusting in 1993. They mine about 30,000 TPY (27,000 MTPY) of limestone from the Tertiary Flagstaff Limestone in eastern Utah County.

6) Rancho Equipment Services mines limestone from the Ordovician Pogonip Group in central Juab County for flue-gas desulfurization at the Intermountain Power Project electric plant, near Delta in Millard County, which requires about 200,000 TPY (181,000 MTPY) of limestone.

7) Western Clay Company mines limestone from the Tertiary Flagstaff Limestone of south-central Utah for coal-mine rock dust and for crushed stone.

8) Magcorp uses oolitic sands from the shores of Great Salt Lake as a stack-gas neutralizing agent in their process for making magnesium metal.

9) Cedarstrom Calcite and Clay produces about 8000 TPY (7200 MTPY) of vein calcite from an underground mine in the Mississippian Deseret Limestone and the Humbug Formation west of Utah Lake for poultry grit.

10) The Coval Company produces limestone from the Mississippian Deseret Limestone from the old Larsen Limestone Company property in Utah County for flue-gas desulfurization at the Bonanza power plant near Vernal.

Three of the limestone operations began production since 1993: Cotter Corporation, Emery Industrial Resources, and Rancho Equipment Services.

Sand and gravel

The bulk of sand and gravel production in Utah comes from Pleistocene Lake Bonneville shoreline deposits along the Wasatch Front urban corridor from Provo in the south to Brigham City in the north. There are four major sand- and gravel-bearing Lake Bonneville benches or deposits which mark relatively long-lived, stable shorelines of the lake. The two highest benches, the Bonneville and Provo benches, provide most of the sand and gravel in the state. The Bonneville bench was deposited at an elevation of 5090 feet (1551 m) above mean sea level (Currey and others, 1984), nearly 900 feet (270 m) above the current elevation of Great Salt Lake. The Provo bench was deposited at an elevation of 4740 feet (1440 m) above mean sea level or about 550 feet (170 m) elevation above the current level of Great Salt Lake (Currey and others, 1984).

The locations of the active pits are not shown on **Figure 1** due to the large number of these pits, but a list of the pits and their locations is available through the U.S. Mine Safety and Health Administration office in Denver, Colorado. Most of the production from these pits is for construction sand and gravel, but a few companies also produce small amounts of industrial sand.

Development of new sand and gravel operations in Utah will face many obstacles. The sand and gravel development picture for Washington County illustrates some of the challenges to future development. Washington County is not naturally blessed with abundant sand and gravel; urbanization, environmental concerns, and land use restrictions will make it difficult to insure a future supply of high-quality aggregate in the county (Blackett and Tripp, in press).

Phosphate

Most of the phosphate resource of Utah is contained in Permian rocks; a smaller resource can be found in Mississippian rocks. The phosphatic shales of the intertongued Park City and Phosphoria Formations of Permian age are the source of most of the phosphate production in Utah. The only present commercial operation is S.F. Phosphate Ltd.'s Little Brush Creek mine in northern Uintah County in the Park City Formation. S.F. Phosphate Ltd., a partnership between Farmland Industries of Kansas City, Missouri and J.R. Simplot of Boise, Idaho, recently purchased this operation from Chevron Chemical. S.F. Phosphate Ltd. mines roughly 2.5 million tons (2.3 million mt) of ore annually which is processed into 900,000 TPY (820,000 MTPY) of concentrate (Bon and others, 1996). S.F. Phosphate Ltd. transports its concentrate northward across the Uinta Mountains through a 90-mile (145-km)-long underground slurry pipeline to Rock Springs, Wyoming for treatment with by-product sulfur recovered from natural gas produced in southwestern Wyoming (Woody, 1986).

Mississippian phosphatic shales of the Delle Member of the Deseret Limestone and the Delle Phosphatic Member of the Little Flat Formation contain a smaller resource and have not been exploited.

Crushed stone

Utah has nearly unlimited sources of high-quality rock suitable for crushed stone throughout the state. The importance of this resource will increase as sand and gravel deposits in urban areas are depleted or made inaccessible by residential development. Production of crushed stone for aggregate has created a regulatory debate in Utah. The Monroc, Inc. and Staker Paving Company quarries at Beck's Spur, north of downtown Salt Lake City, produced sand and gravel for many years but recently exhausted their sand and gravel reserves and began crushing bedrock at their pits. Sand and gravel production does not require reclamation permitting in Utah and when Monroc switched to crushing stone they claimed that they were still exempt from reclamation; the State of Utah agreed. This has been very controversial in part due to the visibility of Monroc's pit located within a few miles of the Utah State Capitol and close to major transportation corridors.

Clays

Utah has an assortment of sedimentary and hydrothermal clays including bentonite, bloating clay, common clay, Fuller's earth, halloysite, and kaolinite. The two largest clay producers in Utah are Interstate Brick Company and Interpace Industries, Inc., both of which blend common clays from many quarries to produce a variety of brick. Interstate and Interpace together mined about 300,000 tons (295,000 mt) of common clay in 1995 (Bon and others, 1996), some of which come from a quarry north of Evanston, Wyoming. Redmond Clay and Salt Company, Inc. produces bentonite from southern Sanpete County for use in construction and in well-drilling fluids. Western Clay Company produces bentonite from northern Sevier County and western Emery County for construction use. ECDC Environmental LC produced and stockpiled large quantities of clay from northern Emery County in 1994 but did no mining in 1995. ECDC mines clay for construction of waste-containment cells at their massive landfill site in eastern Carbon County. Grand County mines bentonite from northern San Juan County to line water canals. The Utelite Corporation mines the organic-rich Cretaceous Mancos Shale, which is expanded in four rotary kilns into a lightweight aggregate which is used as aggregate in lightweight concrete and as aggregate in concrete roofing tiles.

Gypsum

Utah has one of the largest gypsum resources in the United States. Withington (1964) estimated reserves of 2 billion tons (1.8 billion mt) averaging more than 85% gypsum in beds a minimum of 4 feet (1.2 m) thick

within 30 feet (9 m) of the ground surface. Whereas numerous geological formations contain gypsum, the Pennsylvanian Paradox Formation of the Hermosa Group, the Jurassic Arapien Shale, the Jurassic Summerville Formation, and the Jurassic Carmel Formation contain most of the resource. Seven companies produced roughly 300,000 tons (270,000 mt) of gypsum in both 1994 and 1995 (Bon and others, 1996). Most of the production was used for manufacture of wallboard but some was used for Portland cement retarder, agricultural gypsum, wallboard joint compound, plaster, and fireproof cores for doors. U.S. Gypsum Company mines Jurassic Arapien Shale gypsum near their Sigurd plant for the manufacture of wallboard, plaster, and wallboard joint compound. U.S. Gypsum is also in the process of permitting a new gypsum mine in Emery County. T.J. Peck and Sons, Inc. mines gypsum from the Arapien Shale near Nephi for export to a cement plant at Inkom, Idaho and for the Ash Grove Cement Plant in Juab County. Georgia Pacific Corporation mines gypsum from their Emery County quarry to supply their Sigurd plant which produces wallboard, plaster, and fireproof cores of doors. H.E. Davis mines gypsum from the Arapien Shale near Levan for shipment to the Holnam, Inc. cement plant in Morgan County. Robert Steele and Jack Dansie also mine gypsum from the Arapien Shale at Levan. D.K. Gypsum Industries mines Jurassic Carmel Formation gypsum from their DKG quarry in Emery County for agricultural gypsum which is shipped to California. Western Clay mines Jurassic Carmel Formation gypsum from western Emery County for sale to Georgia Pacific.

Building stone

Utah companies quarry quartzite, sandstone, and limestone (Tripp, 1993). Companies produce quartzite flagstone and ashlar in Box Elder County from the Proterozoic Elba Quartzite and Quartzite of Yost and the Cambrian Clarks Basin Quartzite (Tripp, 1994). Companies operating in this area include Northern Stone Supply, Utah Building Stone, State Stone, and American Stone. Quartzite from Box Elder County has been shipped since the 1950s but the market has really begun to grow recently with increased national and international demand, the consolidation of small "mom and pop" operations by larger, better capitalized firms, the strong Western economy, and improved technology for producing tiles from quartzite blocks. Sandstone is quarried from the Jurassic Nugget Sandstone in Summit County for local consumption in Park City and Salt Lake City and shipment to the west coast by American Stone and Alan Hansen. Red and tan sandstone blocks are extracted from the Triassic Moenkopi Formation in Wayne County by Hansen Stone and State Stone. Gray sandstone is extracted from southern Duchesne County by Hansen Stone. Small amounts of white, oolitic Tertiary Green River Formation limestone is quarried in Sanpete County by State Stone.

Gemstones

Utah has long produced decorative and semi-precious gemstones on a small scale; Utah red beryl, however, has potential to be developed into a more significant commodity. Kennecott Exploration currently has an option on the Violet Claims which produce a very unusual red beryl of gem quality which has been sold either unprepared or faceted by a Millard County family. Kennecott is currently (1996) drilling the property. The beryl occurs in clay along fractures in devitrified topaz rhyolite of Tertiary age. The reason that the beryl may be found at the Violet Claims and not in similar rocks elsewhere is due to the incursion of abundant surface water into still-hot rhyolite (Christiansen and others, 1996).

Miscellaneous commodities

Other industrial rock and mineral commodities that are lower in total dollar value than the nine commodities previously discussed but which are still important include silica, perlite, and scoria. Silica is produced by A.P. Green from Tooele County for refractory use. Silica is also mined by McFarland and Hullinger in northeastern Tooele County for sale to Kennecott as a copper-smelting flux. A small amount of perlite is mined by Monitor Minerals from the recently reactivated Pearl Queen mine in Beaver County. Scoria is mined from Beaver County by Red Dome Company primarily for decorative landscape granules, but a small amount is also sold for dimension stone.

Commodities present in significant quantities in Utah but not currently in production include: barite, diatomaceous earth, magnesite, fluorspar, pumice, wollastonite, and zeolites (Mayes and Tripp, 1991).

Acknowledgments

The Utah Division of Oil, Gas and Mining provided essential information for this report on current activities of industrial mineral companies. Reviews by J. Wallace Gwynn and Michael Hylland improved the quality of this paper.

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Development of the Kelsey Lake diamond mine, Colorado

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Abstract

The Kelsey Lake kimberlite cluster of the Colorado-Wyoming State Line district has been developed as the first United States diamond mine. Study, evaluation, and development have progressed since their discovery in 1987. Evaluation stage bulk sampling has produced gem-quality diamonds to 14.2 carats (2.94g) in size.

The 30 or so State Line district kimberlites were intruded at approximately 390 Ma into Proterozoic crystalline rocks, presumed to be cored by Archean basement of the Wyoming province. Essentially all contain diamonds, but apparently only the Kelsey Lake pipes are economic at this time.

The Kelsey Lake kimberlites contain predominantly diatreme facies kimberlite, locally occurring with minor hypabyssal and crater facies rocks. Alteration and weathering of the kimberlites is extensive. Petrologic study indicates an underlying lithosphere which has a largely Proterozoic signature and shows evidence of deep melt and metasomatic events.

Development includes a 210 tph (tons per hour) (190 mtph or metric tons per hour) process plant and ancillary water and tailings systems. Two pits have been developed on a diamond-bearing kimberlite resource of 18.7 million tons (17 million tonnes or metric tons) to a depth of 328 feet (100 meters). The mine is currently operating at the nominal rate of 500,000 tpa (tons per year), with future expansions being considered to take the mine to greater than one million tpa. Existing mining permits will be upgraded to allow for the expansion. The environmentally benign nature of diamond mining and processing has facilitated a relatively simple permitting process.

The ore processing comprises a liberation stage, a gravity concentration stage using rotary pans, and a recovery stage using grease tables. The final product has been sold as Colorado Diamond™ *for the first time in 1996.*

Introduction

The Kelsey Lake kimberlites, a cluster of eight diamondiferous pipes, occur in the northern State Line kimberlite district of the Colorado/Wyoming diamond province (Coopersmith 1991, 1993) (**Figure 1**). This location is near the southern margin of the Wyoming Archean province. Two relatively large pipes and a few smaller bodies define the cluster (**Figure 2**). Kelsey Lake kimberlites have been developed as the first diamond mine in the United States.

Current development has concentrated on the two largest pipes. These have undergone a bulk sample evaluation after a geological study (from 1987 on) identified the presence of small diamonds and a mineral chemistry study indicated favorable conditions for the occurrence of diamonds. An 11-tons-per-hour (tph) (10-tonnes-per-hour or mtph) rotary pan plant was established on the site in 1992 for the processing of several thousand tons of bulk sample material. Indicative grades and gem-quality diamonds to 14.2 carats in size were recovered. Early in 1995, it was decided to progress the evaluation program to commercial mining and production commenced in early 1996.

A 210-tph (190-mtph) rotary pan plant has been built. This includes a new water supply system and tailings disposal. Design, fabrication, and construction was completed in-house, with engineering assistance by Lyntek of Denver.

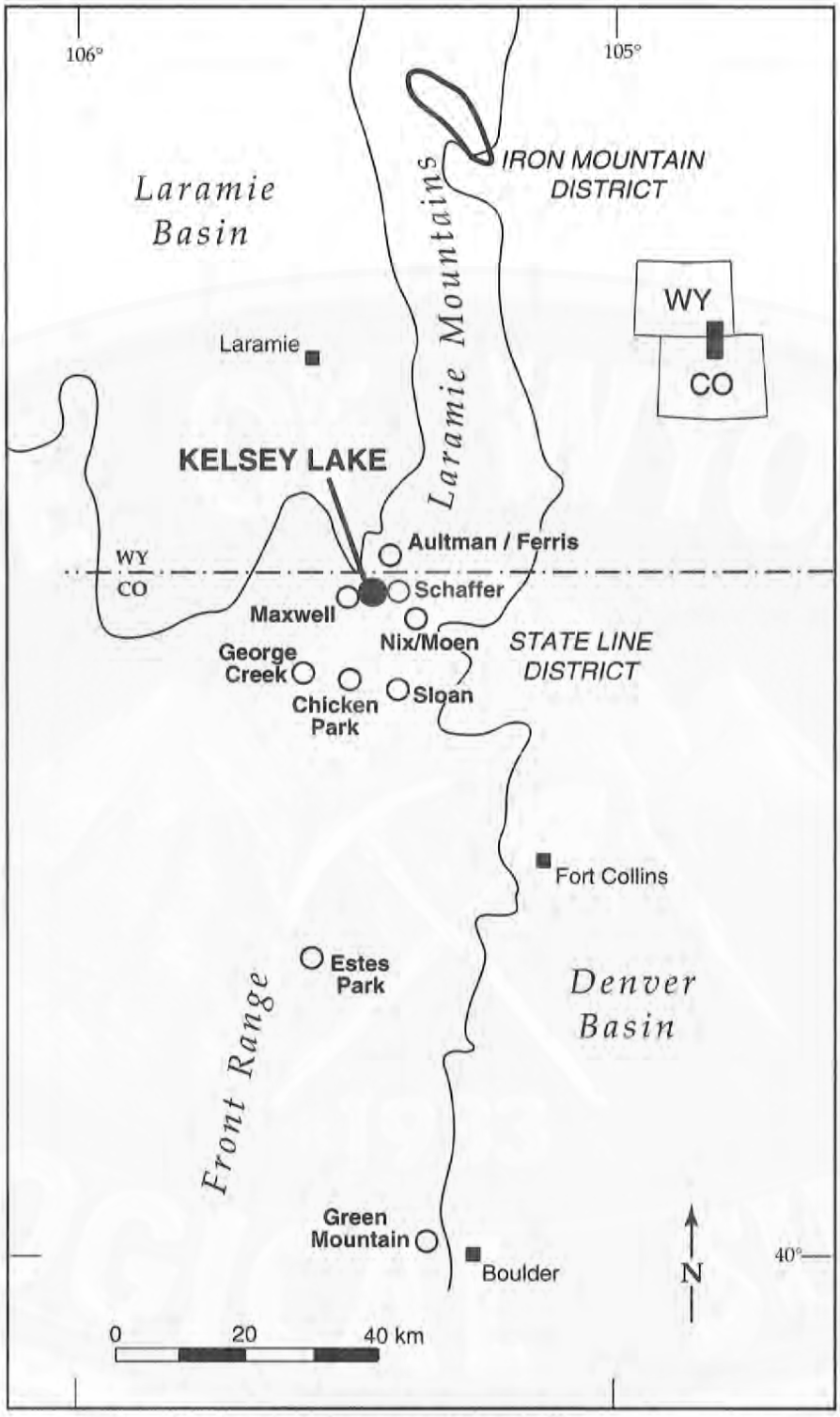


Figure 1. Map of the State Line district of the Colorado-Wyoming diamond province.

Existing mining permits were amended to incorporate the increased scale and new plant facility. Permits for this project, which straddles the Colorado/Wyoming state line, include the following: Colorado mining permit, Wyoming dozing permit, Corps of Engineers Section 404, Colorado National Pollution Discharge Elimination System, Wyoming National Pollution Discharge Elimination System, Storm Water Management Plan, Air Pollution Emission Notice, various Colorado and Wyoming State Engineer and well permits, Substitute Water Supply Plan, Larimer County Special Use Review, and U.S. Forest Service road use permits. Of particular sensitivity is the temporary destruction of a wetlands area, which has been ad-

dressed through the Clean Water Act Section 404 permit administered by the Corps of Engineers. This use will be mitigated by creation of new, self-sustaining, wetlands within the immediate drainage system. A comprehensive and fully bonded reclamation plan is incorporated in the permits.

Diamond mining and ore processing is environmentally benign. No chemicals are used in the processing, which is basically a water wash plant. The kimberlite ore is quite inert, as is the granite wall rock. The exposure of the ore body, the subsequent ore processing, and waste rock and tailings containments pose no threat of release, leaching, or dispersion of any detrimental elements. There is no acid-forming waste, and even a prolonged acid rain environment will not produce a harmful leachate. Slimes containment and clarification ponds will prevent unacceptable levels of suspended solids in water discharge. The isolated restricted access and non-visible nature of the mine and processing has led to rather straightforward permitting.

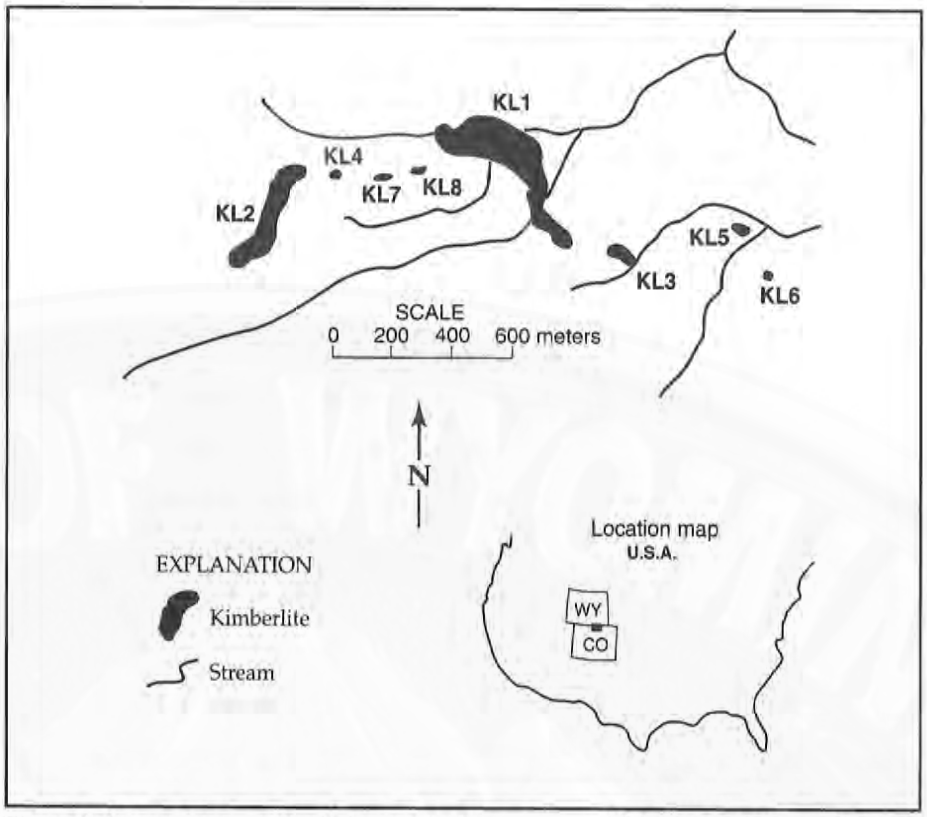


Figure 2. Detailed map of the Kelsey Lake kimberlites and locations of diamondiferous pipes.

Geology

The State Line district is located in the northern Front Range, which is cored by crystalline rocks of a Proterozoic mobile belt (McCallum, 1991). An older (1.8 to 1.7 Ga) metamorphic terrane is intruded by granitic rocks (1.4 Ga). The metamorphic rocks are largely felsic to mafic gneisses and schists. The granitic rocks are comprised of the 1.4 Ga Sherman batholiths and the slightly younger Log Cabin batholith. Dikes of varying composition reflect late stage batholith-related magmatic activity.

Post-Precambrian rocks are restricted to the flanks of the Front Range and consist of Pennsylvanian to Cretaceous sedimentary units. The principal structural features of the Front Range were developed during the Laramide Orogeny, although many are suspected to have a Precambrian ancestry (McCallum, 1991). Although pre-Pennsylvanian sedimentary rocks are presently absent in the region, xenoliths and large blocks of Cambrian, Ordivician, and Silurian sedimentary rocks (limestone, dolomite, conglomerate, sandstone, etc.) do occur in the kimberlite diatremes. Therefore, emplacement of the diatremes (late Devonian) predated an extensive erosional event that removed nearly all early Paleozoic rocks of the region.

Approximately 30 occurrences of kimberlite are known in the district (Smith and others, 1979). These pipes include larger, upper level diatreme facies such as at Sloan and Kelsey Lake; lower level diatreme to root zone intrusions, such as at Nix and Moen; and hypabyssal dikes such as George Creek. Various kimberlites in the district have been evaluated for diamonds, with grade results ranging from zero to close to one hundred carats per hundred tons. Essentially all occurrences of kimberlite in the district contain some diamond, although most are not commercial (McCallum and Waldman, 1991).

The Kelsey Lake kimberlites form a small group of irregular-shaped pipes and fissures (**Figure 2**). The pipes consist largely of multiphase blows of diatreme facies kimberlite with extensions of hypabyssal-facies kimberlite. The kimberlite is altered and deeply weath-

ered. The two main pipes combine to over 20 acres (8.1 hectares) in surface area, ranking them among the largest in the district. The kimberlites intrude 1.4 Ga Proterozoic granitic rocks of the Sherman and Log Cabin batholiths. Shape is largely controlled by jointing in the country rock. The kimberlite contains sedimentary xenoliths of Cambrian to Silurian age. An apparent Devonian age of the Kelsey Lake kimberlite occurrences is in good agreement with Early Devonian (390 Ma) isotopic ages of other kimberlites in the district (Smith, 1983).

The KL-1 pipe consists of multiple phases of diatreme facies and possible crater facies kimberlite. Pellet-rich tuffisitic (volcaniclastic) serpentine kimberlite breccia occupies the southern fissure appendage of the main pipe. This is extremely rich in mantle xenocrysts, megacrysts, and xenoliths. The southern portion of the pipe contains pellet and autolith-rich layered pyroclastic kimberlite, generally of a sandy tuff. The quantity of rounded quartz grains and sedimentary xenolith material varies between and within layers. Large [up to 10-foot (3-meter)-diameter] sedimentary blocks are present. This zone has a moderate amount of mantle xenocrystal and xenolithic material. The main pipe, which comprises the northern part of the body contains tuffisitic (volcaniclastic) serpentine kimberlite breccia, with a moderate amount of mantle component, and locally abundant large xenolithic blocks.

The KL-2 pipe is a multiphase diatreme facies complex with at least four distinct phases of tuffistic (volcaniclastic) serpentine kimberlite breccia. These are locally rich in mantle xenoliths and xenocrysts. Sedimentary xenolith content, size, and preservation varies considerably between phases. Alteration and carbonatization is also variable. Locally abundant crustal xenoliths exceed one meter in size. Occurrence of a contact breccia is indicated.

Late stage alteration and surface weathering has destroyed much of the primary mineral assemblage of both bodies of kimberlite. Ubiquitous pheonocrystal and macrocrystal olivine is totally replaced by serpophitic or lizardite serpentine, and commonly is carbonitized. Groundmass diopside, perovskite, apatite, and opaque minerals generally are preserved. Groundmass serpentine and calcite may be either primary or secondary. Macrocrysts and xenocrysts of Mg-ilmenite, pyrope garnet, spinels, phlogopite, and chromian diopside are generally preserved, as are megacrysts of Mg-ilmenite, Cr-poor titanian pyrope, and subcalcic diopside. Groundmass in autoliths and pellets is relatively fresh, and generally is phlogopite rich. Petrographically these rocks are Group 1 kimberlites. Xenolithic material in the pipes includes upper mantle garnet periodite, spinel periodite, eclogite, lower crustal granulite, and crustal granitic, metamorphic, and sedimentary rocks. Content and size ranges of xenolithic and megacrystal material varies widely within the pipes.

Trace elements in the Kelsey Lake garnets and chromites have been analyzed using the proton microprobe at CSIRO (Commonwealth Scientific and Industrial Research Organization), Syndey, Australia (Coopersmith and others, 1995). The garnet and spinel data has defined a geotherm corresponding to a 35 mW/m^2 conductive model to about 40 to 50 Kb. Above this the geotherm jumps to 40 mW/m^2, showing a deviation from the conductive model above 1100°C. It appears that the mantle was strongly heated at depths below about 109 miles (175 km), and this raised the geotherm in the lower part of the lithosphere.

The lithosphere, as defined by the presence of depleted rocks, probably did not extend deeper than 109 miles (175 km) at the time of kimberlite intrusion. The upper mantle in this region contains only 5 to 10% harzburgite and this is Ca-rich. This is more typical of a Proterozoic lithosphere than an Archean one. This probably reflects the craton-margin settling, where a significant thickness of Proterozoic lithosphere has replaced the Archean rocks at depth. Both potassic and high-temperature melt-related metasomatism are noted at depth.

Evaluation and development

Initial evaluation included delineation and mapping of the kimberlite deposits. Detailed petrographic study and mineral chemistry indicated a strong likelihood that the deposits would host important diamond mineralization. Small diamonds (0.01 to 0.20 carat) were recovered from small surface samples.

In 1990, approximately 1100 tons (1000 tonnes) were processed through a nearby contract diamond processing plant. Recovery at this stage included gem diamonds in excess of 2 carats in size.

In 1992, an 11-tph (10-mtph) test plant was constructed on site. This plant utilizes a rotary diamond pan for concentration and a vibrating grease table for recovery. Approximately 6600 tons (6000 tonnes) were processed through this plant. The resulting diamond recovery included gem diamonds of 6.2 and 14.2 carats.

In 1995, it was decided to commence trial mining at the 275,000-tpy (250,000-mtpy) rate. Permits have been upgraded and amended to accommodate the increased scale of the operation. Process and plant design and construction has been completed. Site development was completed in the fall of 1995, whereupon mining commenced.

The ore processing facilities comprise three sections: liberation, concentration, and recovery. Gravity process is utilized for the concentration, as diamond reports to the heavy mineral concentrate. The liberation stage is critical to disaggregate the ore and release the heavy mineral grains, which include the diamonds. Liberation also completely breaks up the clay, which is critical to the process. Crushing and scrubbing units are utilized in the liberation stage.

The concentration stage uses four rotary diamond pans, each fourteen feet (4.27 m) in diameter. A gravity separation is accomplished using the naturally occurring finely ground up rock in suspension to act as the dense media. With the disaggregated ore in a slurry with the natural dense media, the heavy minerals drop out into a concentrate. Carefully controlled mass balance and pulp density between ore, slurry, and water maintain the desired density separation.

The recovery stage utilizes physical properties specific to the diamond. Diamond is hydrophobic, or nonwetable. The heavy mineral concentrate is passed with water over a vibrating grease table. This is coated with a thin layer of specially formulated "collector grease." As most of the concentrate has a thin film of water adhering to it, it passes over the grease table. The diamond, remaining dry, sticks to the collector grease. This method is extremely efficient and produces a very clean diamond concentrate for final hand sorting.

Commercial production commenced in early 1996. Consumers were able to purchase *Colorado Diamonds*™ for the first time in 1996.

Acknowledgments

This project would not have come to fruition without the continued support of the founding management of Redaurum Ltd., and its predecessor companies. Here, thanks go to R. Baxter-Brown, A.R. Hamilton, and C.E.E. Devenish. Thanks are also extended for permission to present this work. The exploration and development of this project was greatly facilitated by the local staff, permitting consultants, and engineering consultants. Also, deep appreciation goes to M.E. McCallum for his pioneering work in the district. Finally, thanks to all who are buying or will be buying Colorado diamonds in the future.

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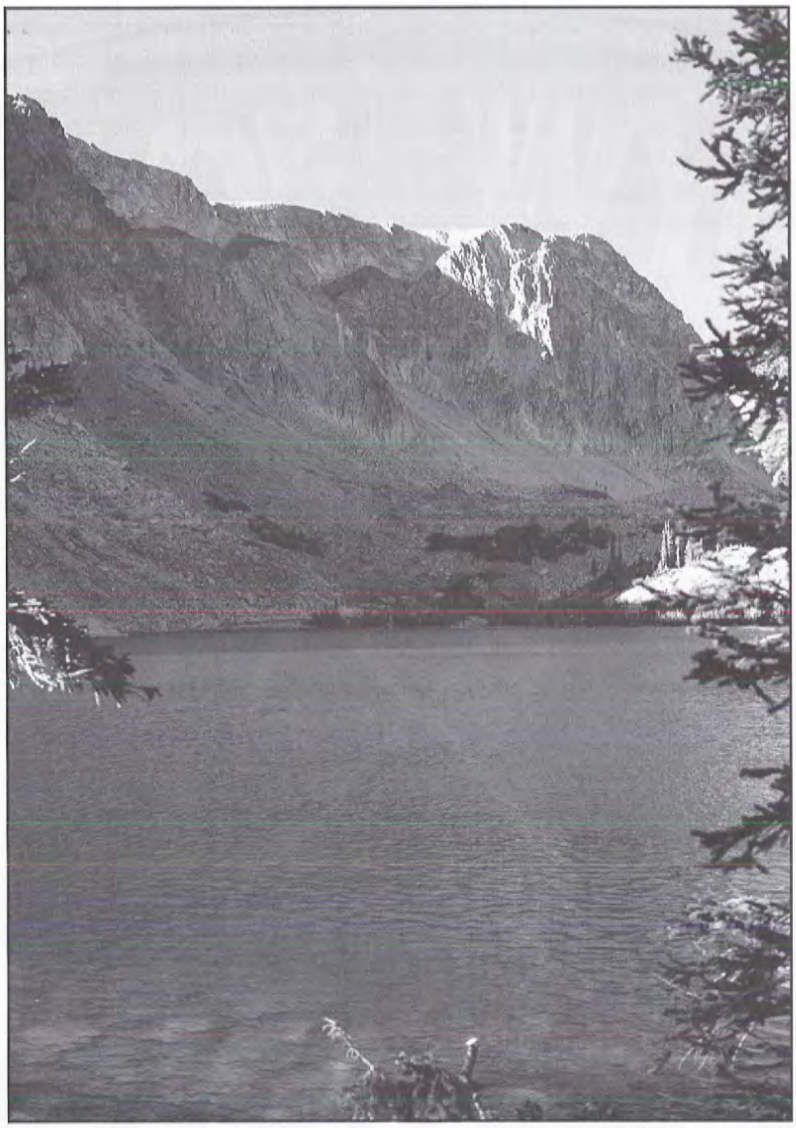
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The eastern front of the Snowy Range as viewed from Lake Marie. This is in Medicine Bow National Forest about 40 miles west of Laramie on the Snowy Range Scenic Highway. Notice the dark dikes that cut vertical beds of highly resistant white quartzite. (Photograph by R.W. Jones.)

A study of quartz extracted from granite

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Abstract

Mineral Technology Corporation (MinTec) produces and sells high purity quartz sands and powders for use by the sophisticated glass and fused quartz industries worldwide. Using a new and proprietary process, MinTec also recycles fused quartz cullet and waste quartz sands generated by its customers during the manufacture of quartz tubing, boules, and crucibles for the semiconductor industry.

Historically, our industry has produced high purity fusion grade quartz from naturally occurring quartz ore deposits. However, as the fused quartz industry continues to impose more stringent specifications, products generated from such quartz deposits have become more and more marginal with respect to quality. Based upon today's requirements few, if any, naturally occurring quartz ore supplies exist in the United States or elsewhere that can meet today's chemical and fusion specifications. It therefore became necessary to identify and examine alternative quartz ore sources.

Approximately two and a half years ago, MinTec discovered that the quartz contained within a certain granite located in the Western United States offered the possibility of becoming a highly desirable ore source. MinTec therefore embarked upon an extensive and possibly unique study of granite deposits primarily located within, but not limited to, eight Western states. Since, MinTec has sampled and tested the quartz separated from approximately two thousand individual granite deposits to learn that most granite yields quartz with one or more undesirable characteristics. There are exceptions, however, and these are the deposits that are of interest to MinTec.

Certain details, primarily pertaining to quartz and the processing thereof, are classified by MinTec as confidential and cannot be published at this time. Other interesting aspects, such as general geology, mineralogy, how each aspect apparently relates to the quality of the quartz, geographical areas of interest, and other findings are discussed in detail.

Introduction

High purity quartz is primarily used by the sophisticated glass and fused quartz industries to manufacture laser glass, fiber optics, photovoltaic cells, wave guides, and other specialty products for the electronics trade. It is also exclusively used to manufacture fused quartz products such as crucibles, tubing, boules, and other products for use by the semiconductor industry. The purity, melting, and fusion characteristics of the quartz used to manufacture these products is critical to the quality of the final glass or fused quartz produced by the manufacturer. Further, as these industries continue to expand technology, the necessity for better quality quartz becomes greater and specifications become more and more stringent.

For glass production wherein the chemical levels of such elements as Al, Na, and K are less critical and transition elements such as Fe and Ti are the major criteria, quartz produced from pegmatites or hydrothermal deposits can now, and probably always will, meet the required specifications. However, the glass trade is but a small segment of the marketplace.

In the production of fused quartz, however, chemical purity and fusion characteristics are of equal importance. Such chemical specifications currently call for Al levels from 4 to 10 ppm with all other elements ideally well below 1 ppm and certain elements, such as boron, becoming critical if they are present in excess of 20 ppb. In addition, the quartz must fuse completely bubble-free. This means that the quartz should not contain any significant amount of gases or hydroxyls within the lattice of the crystalline structure. Based upon these specifications, MinTec's studies have indi-

cated that it is nearly impossible to produce quartz derived from a pegmatite or hydrothermal deposit that will meet the specifications of today's fused quartz industry; our studies further reveal that if quartz products acceptable to this industry are to be produced, such production must be generated from homogeneous granite deposits.

Quartz testing procedures

Testing a quartz sample from a given deposit to learn if it is satisfactory for use by the fused quartz industry requires much more than performing a simple chemical analysis. The sample must first be prepared using a beneficiation process designed to simulate actual plant conditions. When testing a quartz sample from a pegmatite or hydrothermal deposit the sample is first ground into a -50 mesh sand and thereafter specialized magnetic separation and chemical beneficiation is applied. A microscopic and chemical analysis is then performed. Provided the results of these tests show promise, the material is then fused into a small quartz boule for further study. To test the quartz contained within a granite sample, the material is first ground to a -50 mesh sand and then subjected to a five-step flotation process designed to recover the quartz segment from the sample. The separated quartz is then further beneficiated, microscopically and chemically tested, and then fused in a manner similar to the process described above.

Quartz from pegmatite and hydrothermal deposits

Over the past several years, in an effort to identify acceptable ore sources for use by the fused quartz industry, MinTec has studied many pegmatite and hydrothermal deposits with only mixed success. The company has not only studied quartz from deposits throughout the United States, but from several foreign countries as well. Although a number of deposits were found to offer possibilities, final results indicated that only marginal quartz products could be produced from these deposits.

MinTec's studies of pegmatite deposits have included zoned, layered, and homogeneous pegmatites. Obviously our conclusions and chemical evaluations have been focused upon the quartz segments only. Very little attention was given to other minerals present, other than the effect such minerals may or may not have had upon the quality of the quartz within a given deposit.

Recognizing that the geology and the mineralogy contained within most pegmatites are generally considered to be somewhat unpredictable, we are continuing to test all deposits that may appear to offer potential. However, after testing the quality of quartz from many hundreds of deposits over a broad geographical area, we have made several observations that thus far appear to be consistent: (1) The chemical quality of the quartz contained within an individual deposit does not significantly vary throughout the deposit unless a drastic visual difference in the quartz is apparent. (2) The chemical quality of the quartz located adjacent to or near another deposit may, and probably does, have completely different chemistry. (3) The quartz contained within most pegmatite deposits normally possesses very poor fusion characteristics. Should it be determined that quartz from a pegmatite deposit does fuse satisfactorily, this quality is generally not consistent within the deposit and production from such a deposit will require extensive hand sorting to maintain acceptable fusion qualities. This type of hand sorting is being done in Brazil, as an example, and routinely results in high cost ore. (4) The presence of excessive amounts of lithium minerals, tourmaline or beryl, while they may not actually penetrate into the quartz itself, generally denotes high alkali and Al levels within the quartz. (5) Excessive amounts of biotite mica and/or muscovite mica within a deposit generally indicates that the quartz may have somewhat improved chemistry.

Our studies further indicate that quartz contained within a deposit of hydrothermal origin generally has better chemistry than quartz derived from a pegmatite deposit. The quartz from a hydrothermal deposit normally has poor fusion characteristics but there are exceptions and, if so, the fusion quality may be consistent throughout the entire deposit.

The chemistry of quartz from a pegmatite deposit almost without exception has high Li content (about 2 ppm or more) and this results in a correspondingly high Al content of about 20 ppm to much higher, depending generally upon the levels of Li present within the quartz. The combined Na and K within a better quality quartz derived from a pegmatite deposit generally is in excess of 4 ppm and from most deposits it will be much higher. The Fe content can normally be reduced to very low levels, 2 ppm or less, but the Ti and Zr may be at any level. Also we have not found any indicators that can be used to estimate these elements. Only a chemical analysis will reveal this

information and if excessive amounts of Ti are present, it generally cannot be removed. Conversely, Zr can be removed provided it is in a certain mineral form. Chemical analyses covering several of the better grades of quartz derived from pegmatite deposits are given in **Table 1.**

The chemistry of quartz extracted from a hydrothermal deposit, on average, is similar to the quartz from a better pegmatite deposit; however, there are exceptions. As an example, the Li can be very low (less than 1 ppm), and the Al can be 4 to 8 ppm with all other elements being in the 1 to 2 ppm range. The primary problem with quartz from a hydrothermal deposit is that generally it contains varying amounts of metallic minerals such as copper and pyrites. Often these elements are microscopic and in such small quantities that alarming chemistry is not detected. These metal elements come back to haunt the fusion process, however, imparting small bubble clusters, pimples, and small black specks within the fused quartz glass. The chemical analyses of quartz from hydrothermal deposits are given in **Table 1.**

In summary, quartz from both pegmatite and hydrothermal deposits offers very little potential as an ongoing and economical ore source for the production of high-purity quartz products for use by the fused-quartz industry.

Table 1. Comparison of chemical analyses of quartz from pegmatite and hydrothermal deposits.

Description		Chemical analysis in ppm										
(Better grades)	**Location**	**Al**	**Ca**	**Fe**	**K**	**Li**	**Mg**	**Mn**	**Na**	**Ti**	**Zr**	**Fusion**
Pegmatite	Black Hills, SD	20.0	1.5	.20	.5	2.0	.10	.01	3.3	.20	.02	Poor
Pegmatite	Black Hills, SD	22.1	1.4	.10	.9	2.4	.09	.02	1.9	.30	.06	Poor
Pegmatite	Glenrock, WY	53.8	3.6	.70	9.4	3.5	.10	.10	18.4	1.40	.60	Poor
Pegmatite	Dubois, WY	31.6	1.4	.90	6.1	2.6	.10	.10	3.6	2.00	1.40	Poor
Pegmatite	Brazil	9.0	1.3	.30	.6	1.0	.10	.01	2.1	.24	.03	Good
Hydrothermal	Black Hills, SD	20.5	1.3	1.00	4.7	.5	.80	.20	5.5	.95	.04	Poor
Hydrothermal	Wheatland, WY	18.2	1.6	.80	1.6	2.1	.01	.01	7.8	5.80	1.20	Poor
Hydrothermal	India	5.4	.7	.40	1.0	.1	.20	.10	1.8	1.90	.60	Good-inclusions

Quartz from granite deposits

Granite is composed of a mixture generally consisting of feldspar (60 to 65%), mica (10 to 15%), iron-bearing minerals (1 to 2%) and quartz (20 to 25%). MinTec has discovered that quartz derived from certain granite possesses unusual and often exceptional chemistry and fusion characteristics. Unlike the somewhat unpredictable quality of quartz recovered from pegmatite deposits, our research has determined that quartz derived from granite is generally very consistent. Our study also has revealed that most geologists tend to overstate the percentage of quartz contained within a deposit. On the other hand, MinTec's data is based upon recoverable quartz and may not be representative of the total amount of quartz that is actually present. However, we have learned that it is almost impossible to visually estimate the percentage of quartz contained within a sample as, once liberated, individual feldspar crystals tend to become clear and practically indiscernible from the quartz. It is also important to note that quartz from within a typical granite contains about 15 ppm of Ti (see **Table 2**) and this alone eliminates such quartz from possible use by the fused quartz industry. Therefore, the thrust of MinTec's study has been to identify unusual and unique deposits wherein the Ti levels are found to be 2 ppm or less (see **Table 2**).

The term granite as used in MinTec's study encompasses a broad range of Precambrian rock types often referred to by others as granodiorite, pegmatitic granites, homogeneous pegmatites, gneiss, alaskite, and many other more descriptive definitions. Often the author of an investigative publication will assign a nomenclature to a given mineral occurrence based largely upon crystal size and structure, together with perceived mineral content. However, many such written reports and associated geologic maps were used as valuable and effective tools in assisting MinTec with the extensive field work required by this study.

Table 2. The range of chemical analyses used to establish MinTec's criteria. Analyses in ppm (parts per million).

Description	Al	Ca	Fe	K	Li	Mg	Mn	Na	Ti	Zr	Fusion
Typical granite	8-30	1-3	.2-4	1-5	1-2	0-1	0-1	1-4	15-25	1-15	Fair-good
Typical pegmatite granite	20-50	1-5	.2-4	1-10	2-10	0-2	0-2	2-20	5-25	1-15	Fair-good
Unusual granite	8-20	1-3	.1-1	.5-1	.01-1	0-.1	0-.1	.5-.1	1-5	1-15	Good
Unique granite	3-20	.5-1	.1-1	.5-1	.01-1	0-.1	0-.1	.5-1	.5-.2	1-15	Excellent

To effectively implement this study, it became necessary for MinTec to establish unique criteria that take into consideration crystal size and structure but are primarily based upon the proven chemistry of the quartz contained within a deposit. Once a deposit was identified as meeting these criteria, it was simply defined as typical granite or possibly granite gneiss. If the chemistry of the quartz from a deposit more closely resembled that of a pegmatite quartz, this deposit was referred to as a pegmatitic granite and most likely was not subjected to further investigation. Better grade granite was identified as unusual, exceptional granite was labeled as unique. (See **Table 2**)

MinTec's study has produced some interesting discoveries and, as the study progressed, many of these findings were incorporated into the criteria that were used to identify additional deposits. Discoveries of major importance are described below.

(1) The quartz from granite deposits, almost without exception, possesses good to excellent fusion characteristics.

(2) A granite deposit containing primarily soda feldspars, such as plagioclase or albite, generally yields quartz with low alkali levels and deposits containing abundant potassium feldspars, such as microcline, often yield quartz containing higher alkali levels. The reason for this is thought to be that potash feldspars have a tendency to become intergrown with the quartz crystalline structure and therefore cannot be liberated, whereas soda feldspars do not.

(3) Granite deposits containing high levels of biotite micas and muscovite micas or a combination of both, tend to yield quartz with more desirable chemistry.

(4) The size of crystallization, although a consideration, may not necessarily affect the chemistry of the quartz. Larger crystallization generally proves to constitute a pegmatitic granite containing higher Li levels with correspondingly higher Al, this holds true with very few exceptions. There are some exceptions, however. For example, the Stockade granite deposit, located in the Black Hills of South Dakota, contains some larger crystallization and what appears to be numerous pegmatite intrusions that normally would denote a pegmatitic granite, yet the chemistry is consistent throughout the deposit and indicates it is of granitic origin. The Li content is very low (.03 ppm) but considering other chemistry, the Al content (20 ppm) is much higher than it should be. We have no explanation for this other than this is a unique granite deposit. Interestingly, however, Norton and Redden (1990) classified and mapped this deposit as part of the Harney Peak granite pluton.

(5) If the chemistry of the quartz indicates that a deposit is of granitic origin, the quartz contained within any pegmatite-appearing intrusions generally will prove to have the same chemistry as the main granite deposit.

(6) The presence of tourmaline, beryl, lithium-bearing minerals, and other rare earth elements, individually or in combination, as commonly found in zoned pegmatites, will almost always denote a pegmatitic granite and the chemistry of the quartz contained in such a deposit will likewise prove this to be the case.

(7) The Ti and Zr levels contained within the quartz from a granite deposit has proven to be positively unpredictable and in many cases no founded explanation for the presence of these elements within the crystalline structure of the quartz can be clearly identified. Not until the final chemical analysis will the extent of these elements be learned. Ti can sometimes be slightly reduced using special beneficiation methods, but more often Ti will stubbornly remain. Zr, on the other hand, can generally be totally removed using known processing methods.

(8) Deposits containing excessive biotite mica generally yield quartz with lower boron levels. One explanation for this is that during cooling and crystallization of the magma, the boron possibly gravitates into the biotite mica and therefore does not combine with the quartz. This supposition, in our opinion, has not been proven.

Granitic areas investigated

Scope of investigation

MinTec's study of granite ore sources has encompassed a broad geographic area. Approximately 2000 samples were collected and evaluated from South Dakota, Wyoming, Montana, Idaho, Washington, Oregon, Colorado, New Mexico, Arizona, and California, as well as from other U.S. locations and foreign countries. Most deposits investigated proved to be typical granite and while such deposits offer scientific value they are of little economic interest to MinTec. Our study has, however, enabled us to identify a small number of unique granite deposits located in the Black Hills of South Dakota and specific locations in Wyoming that offer potential.

South Dakota

South Dakota has only limited areas containing Precambrian granitic rocks of significant proportions, these being primarily the Black Hills, located in the southwestern part of the state; Milbank, located in the southeastern part of the state; and a very small area in the central part of the state. With the exception of the Black Hills, all other areas within the state contain typical to substandard granite. Many deposits located in the Black Hills, primarily within the Harney Peak granite pluton, yield quartz with less Ti than typical granite and a few have been identified as unique granite deposits having exceptional chemistry overall. The chemical analyses for some of these deposits are in **Table 3**.

For many years the geology within the Black Hills has been studied by many capable geologists and a raft of publications, including geologic maps, are available. Many of these publications were used to assist our study, including, but not limited to, the following: Van Hise (1896); Norton and Redden (1990); Duke and others (1988, 1990); Redden and others (1985); Shearer and others (1987). Even with the assistance provided by such geologic maps and publications (which obviously were not focused upon quartz mineralization), we have been unable to establish any clear cut correlation between a specific geological event or rock type and the quartz found in the granite. Therefore, we could not establish guidelines for identifying granite containing quartz with potential for development. The unique or unusual deposits that we have been able to identify are randomly situated throughout the entire southern Black Hills area and are most often surrounded by other deposits containing typical or pegmatitic granite and, in one case, zoned pegmatites.

Wyoming

Wyoming has many areas that contain Precambrian granitic rock types containing granite deposits with potential. For this reason, a great amount of field work and testing was required to eliminate areas that did not appear to offer potential and to establish specific areas of primary interest. After this preliminary reconnaissance had been completed, we began concentrating our efforts within the areas identified as having the greatest potential.

The general areas that were investigated and thus far offer only limited potential are as follows: (1) The Black Hills of Wyoming, located in the northeastern corner of the state, contain essentially deposits of pegmatitic and typical granite. (2) The Bighorn Mountains, covering an area from about Sheridan on the north to just south of Buffalo, contains primarily typical granite. (3) The Copper Mountain area just north of Shoshoni contains primarily pegmatitic granite together with a few typical granites. (4) The Du Noir area north of Dubois proved very disappointing and primarily contains deposits ranging from pegmatitic granite to substandard typical granite. (5) The Granite Mountains from Jeffrey City to Alcova contain primarily typical granite. (6) The Reno Hills area south of Casper and Glenrock, including Esterbrook Mountain, contains pegmatitic and typical granite. (7) The Laramie Range east of Laramie contains primarily anorthosite, together with some typical granite occurrences. (8) The Snowy Range near Albany to east of Riverside contains typical granite. (9) The Green and Ferris Mountains south of Jeffrey City and Muddy Gap proved to contain typical granite.

After our preliminary investigation certain general areas appeared to offer potential. Therefore we began and are continuing more extensive studies within the following areas.

(1) The Wind River Range from about Lander to Pinedale, including the South Pass area generally contains typical granite, with the exception of certain deposits situated near Atlantic City. While we were unable to identify an individual deposit containing quartz with ideal chemistry, we did find deposits with exceptionally low levels of Ti. As this is an unusual occurrence in Wyoming, this small area is scheduled for further investigation.

(2) We were able to identify one unique deposit and several other deposits containing quartz with low Ti levels in the Laramie Mountains west of Wheatland, including the Cooney Hills, and to just

Table 3. Comparison of chemical analyses of quartz from granite.

Description (Better grades)	Location	Chemical analysis in ppm Al	Ca	Fe	K	Li	Mg	Mn	Na	Ti	Zr	Fusion
Pegmatitic granite	Black Hills, SD	32.8	2.9	.4	6.2	3.10	.20	.20	4.9	5.8	1.30	Fair
Typical granite	Black Hills, SD	23.5	1.6	.5	4.6	2.10	.30	.10	1.9	15.2	3.10	Fair
Unusual granite	Black Hills, SD	8.3	.6	.3	.5	.14	.01	.01	1.5	4.4	3.20	Good
Unique granite	Black Hills, SD	19.9	.5	.2	.5	.10	.10	.02	.9	.5	.02	Excellent
Typical granite	Bighorn Mts., WY	16.2	1.2	1.5	2.8	1.20	.20	.10	3.9	16.2	1.90	Fair
Unusual granite	Laramie Peak, WY	9.0	.1	1.4	3.2	.02	.10	.10	1.8	5.5	1.10	Good
Unique granite	Laramie Peak, WY	11.4	.8	.9	1.0	.07	.20	.10	.8	1.8	.90	Excellent
Unusual granite	Encampment, WY	7.1	1.0	.5	1.2	.01	.30	.01	.8	2.9	.90	Excellent
Unique granite	Encampment, WY	4.2	.5	.4	.5	.01	.20	.10	.8	1.5	.93	Excellent

north of Laramie Peak. Our investigation of this area is continuing.

(3) Within an area from northeast of Encampment and southeast of Riverside to about Blackhall Mountain we have identified several unique granite and granite gneiss deposits. One of these deposits has been extensively tested and it has been determined to consistently yield quartz that will meet certain specifications required by the fused quartz industry. Production from this deposit is planned to begin in 1996. The chemical analyses for some of these deposits identified as having potential are in **Table 3**.

Publications and geologic maps used to assist in our study of Wyoming granite include, but are not limited to, the following: Keefer (1957); McCallum and Orback (1968); Hausel (1982); St. Clair and others (1959); Ferris (1964); Love and others (1955); McCallum (1968); and Merry (1963). With the exception of certain geologic maps, most of the publications were obtained after a given deposit of interest had been identified; they were used to facilitate a more thorough survey of a specific area.

Idaho

The batholith granitoid formations are found over a large portion of central Idaho, covering an area essentially from Coeur d'Alene on the north to Boise on the south and extending to the Idaho-Montana border on the east. While the mineralization contained within the batholith occurrences appeared visually attractive, many samples over a broad area have proven these deposits to be typical granite. The quartz consistently contained from 15 to 25 ppm of Ti. Quartz from numerous residual clay deposits originating from the batholith was also tested and while some improvements were noted, the excessive levels of Ti persisted. At this time no further study of these deposits is planned.

Montana

Our study of Montana granite has been primarily confined to specific areas located within the southwestern portion of the state. Numerous samples tested indicate that this area contains typical to substandard granite. The Elkhorn Mountains were extensively studied as it was reported that deposits of alaskite and granodiorite abundantly occurred there (Klepper and others, 1957, 1971). Our study proved that quartz from these areas contained very high levels of Ti and other undesirable elements.

Colorado

Our study in Colorado primarily encompassed the Front Range, essentially from border to border north and south and specific areas near Gunnison and Lake City on the western slope. We also investigated deposits within the Salida area. All of the samples tested proved to be typical or pegmatitic granite and nothing of interest was identified.

New Mexico

In New Mexico our studies were limited to specific areas from about Santa Fe north to the Colorado border. Particular attention was given to the Petaca

and Ojo Caliente area. All of the samples tested proved to contain exceptionally high levels of Ti with other undesirable elements and no deposits of interest were discovered.

Other areas

Our studies within areas not described above were most limited and cannot be considered conclusive even within the areas investigated. However, nothing of interest was identified from any of these locations. In Washington, as an example, we investigated several deposits just west of Moscow, Idaho and a small area southwest of Spokane. In Oregon we investigated a small area near Grants Pass. In Arizona we surveyed and investigated several deposits northeast of Tucson near Mt. Lemmon and in a small area near Prescott. In California we surveyed and tested several areas in the San Gabriel Mountains and small areas both north and south of San Bernadino. We have also tested samples submitted by others from areas within the northeastern and southeastern United States and from most Central and South American countries. In addition, we have tested numerous samples from China, Russia, India, and Portugal.

Ongoing study of quartz ore sources

The fused quartz industry has historically required and is expected to continue demanding purer and purer sands and powders. In addition, as the complexity of specifications continues to expand, the industry is also experiencing enormous growth. This means that the demand for larger quantities of quartz products will become greater and greater. Further, as the various technologies advance, the need for quartz products engineered to meet a specific requirement will become extremely important. To do this will require production from many quartz ore sources as opposed to a single ore source. This is because the industry's ability to beneficiate a single quartz material to meet various specifications has proved to be most limited and in many cases impossible.

It therefore becomes obvious that the need to discover and exploit new quartz ore sources must be ongoing. For these reasons, MinTec anticipates continuing our study to identify granite deposits that offer potential and based upon the data assimilated thus far, we currently plan to concentrate our efforts within the Black Hills of South Dakota and certain areas of Wyoming.

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Ultramafic lamprophyres from Porcupine Dome, east-central Montana, and their potential for diamond

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Abstract

Isolated dikes and diatremes occur in the Froze-to-Death Butte, Geyser Spring, Johnson Ranch, and Gold Butte intrusions of Porcupine Dome in the northern Wyoming craton. Geological, mineralogical, and chemical evidence is used to characterize the least-altered rocks of the intrusions. Although they share some characteristics with kimberlites, the intrusions are best classified as aillikites, a variety of phlogopite-olivine-carbonate ultramafic lamprophyres. Even the best preserved rocks have been altered by carbonatization and oxidation processes. Whole-rock and mineral compositions indicate that the magmas originated in the upper mantle.

The Froze-to-Death Butte and Johnson Ranch intrusions also contain two important kimberlite indicator minerals, pyrope garnets and picroilmenite. The pyropes, recovered from ant hills and crystalline outcrops, range in size from 1 to 4 mm and most have fresh conchoidal fracture surfaces. Other crystals exhibit kelyphitic alteration rims and veins. Two color-compositional groups of pyropes are recognized, red-purple garnets and pink-orange garnets. Red-purple garnets contain, in weight percent (wt. %), 5.7 to 7.0 Cr_2O_3, 4.7 to 7.0 CaO, 18.6 to 21.6 MgO, and 5.9 to 7.9 FeO. The compositions lie within the range of African kimberlite garnets derived from lherzolite sources and are similar to peridotite and megacryst garnets from kimberlites of the Missouri River Breaks diatreme field of north-central Montana. Most red-purple garnets are considered "G9" Cr-pyropes normally associated with diamond-barren lherzolitic sources. However, a few grains are subcalcic and have "G10" compositions, similar to South African garnets derived from diamond-bearing harzburgitic sources.

Pink-orange garnets have (in wt. %) 0 to 0.1 Cr_2O_3, 1.0 to 4.5 CaO, 8.1 to 13.2 MgO, 22.8 to 30.8 FeO, < 0.06 TiO_2, and < 0.07 Na_2O. Chemically they best resemble similarly-colored garnets from the Missouri River Breaks derived from Precambrian schists. Alternatively, the pink-orange garnets from Porcupine Dome may be derived from eclogites.

The picroilmenites are 1 to 5 mm in size and have intermediate levels of MgO (5 to 13 wt. %), low Cr_2O_3 (< 0.7 wt. %), and crystal rims that are enriched in MgO relative to cores. They are similar to picroilmenite macrocrysts from kimberlites.

The geological setting, the mineralogical and chemical compositions, and the presence of Cr-pyropes with "G10" compositions are favorable indicators that the Porcupine Dome magmas were derived from a potentially diamondiferous upper mantle source. However, the picroilmenite compositions and the whole-rock alteration suggests oxidizing conditions in the magmas that would have been unsuitable for the survival of diamonds. Although the Porcupine Dome exposures are probably too small for economic diamond recovery, ultramafic lamprophyres elsewhere are worth examining more closely as a potential host for diamonds.

Introduction

The use of indicator minerals such as pyrope, Cr-diopside, picroilmenite, and chromite to locate kimberlites, lamproites, and, ultimately, diamonds, is well established in diamond exploration. Certain chemical signatures in these minerals are used to provide clues as to whether the original host rocks may carry diamonds. For example, purple peridotitic garnets with low-Ca, Cr-rich compositions are commonly associated with diamond-bearing kimberlites (Meyer, 1968; Meyer and Boyd, 1972) and efforts are made to recover a high proportion of purple garnets in order to increase the chances of finding sub-calcic compositions (e.g., McGee, 1988; Simpson, 1993). Although kimberlites and lamproites are normally the host rocks targeted for diamond exploration, a few microdiamonds have been found in ultramafic lamprophyres (Rock, 1991).

Unusual igneous rocks in the form of dikes, diatremes, and sills occur within the northern Wyoming craton (**Figure 1**), in small fields isolated from major intrusive centers of the well-known central Montana alkalic province (CMAP). In north-central Montana, kimberlites and other ultramafic igneous rocks occur in the Missouri River Breaks area (Hearn, 1968, 1979) and lamproites are found at one locality, the Smoky Butte intrusion (Matson, 1960; Mitchell and others, 1987). The Porcupine Dome and Grassrange areas of central Montana also host igneous intrusions (**Figure 1**). The Winnett sills of Grassrange have been recognized as alnöites for some time (Ross, 1926), but other intrusions in this field and in Porcupine Dome have been investigated only recently. The recognition of kimberlite indicator minerals in the Porcupine Dome intrusions (Doden and Gold, 1993a), the kimberlitic characteristics of the rocks (Doden and Gold, 1993b), and the current interest in diamond exploration in western North America encourages further examination of these intrusions for their economic potential. The goal of this paper is to evaluate the potential for diamond in igneous rocks of Porcupine Dome based on the rock types present and the chemistry of kimberlite indicator minerals. The Grassrange igneous rocks are discussed elsewhere (Doden and Gold, 1993b; Doden and others, 1994; Doden, 1996).

Geological setting

The Porcupine Dome intrusions are located in the northern Wyoming craton (**Figure 1**), an Archean-age segment of continental crust that has experienced a complex tectonic history (e.g., Karlstrom and Hous-



Figure 1. Map of the northern Rocky Mountain region showing the approximate boundary of the Wyoming craton (dashed line) and important lamprophyre, kimberlite, and lamproite localities (circles). Ultramafic lamprophyres occur in Porcupine Dome (see inset for specific localities), Grassrange, and the Missouri River Breaks. Kimberlites occur in the Missouri River Breaks, Iron Mountain and State Line districts, Estes Park, Green Mountain, and Cedar Mountain. Lamproites occur at Smoky Butte, the Leucite Hills, and Moon Canyon/Whites Creek. Information from Hearn (1989b), Doden and Gold (1993b), Mitchell and others (1987), Mitchell and Bergman (1991), and Hausel (1996). Modified from Doden (1996) and Hausel (1996).

ton, 1984; Green and others, 1985; Houston and others, 1993). Deep-seated fracture zones reactivated repeatedly over time (e.g., Brown and Brown, 1987; Shurr and others, 1989) cross the craton and appear to have been the loci for Tertiary igneous activity that formed the CMAP. Most igneous rocks of the CMAP occur in relatively large centers dominated by stocks, laccoliths, sills, and dikes with relatively minor amounts of extrusive rocks. A wide range of rock types exists, including syenites, monzonites, phonolites, shonkinites, and granites, with minor occurrences of latites, diorites, lamprophyres, and carbonatites (Hearn, 1989a; Baker and Berg, 1991). Tertiary igneous rocks also occur as isolated dikes, diatremes, and rare sills within the CMAP, primarily in the Missouri River Breaks, Grassrange, Porcupine Dome, and Smoky Butte localities (**Figure 1**).

The crystalline basement and 2000 to 4000 feet (610 to 1220 m) of Precambrian clastic rocks of the Belt Supergroup (Shepard and Precht, 1989) are overlain by Cambrian through Paleocene shallow shelf and terrestrial sedimentary rocks. Phanerozoic deposition was dominated by sandstone, mudstone, and carbonate sediments that repeatedly covered the Western Interior shelf in broad, onlapping sequences (Maughan, 1993) that now total 5000 to 9000 feet (1525 to 2745 m) in most areas of central Montana (Nelson, 1993). The present erosional surface in central Montana exposes nearly flat-lying, gently folded "middle" to Upper Cretaceous clastic sedimentary rocks.

Porcupine Dome intrusions

Four intrusions are known in or near Porcupine Dome: Froze-to-Death Butte, Geyser Spring, Johnson Ranch, and Gold Butte (**Figure 1**, inset). The orientation of the intrusions and *en echlon* arrangement of some dikes suggest a structurally-controlled emplacement. No reliable age dates have been determined, but Hearn (1989a) estimated an Eocene age based on intrusive relationships in the Johnson Ranch diatreme. A single K-Ar age [78.6 ± 2.7 Ma (million years ago)] obtained from large phlogopite crystals at Froze-to-Death Butte is considered spurious (Marvin and others, 1980); it is possible that the crystals analyzed were xenocrysts (see below).

Froze-to-Death Butte is the erosional remnant of a small (250-foot [76 m]-diameter), subcircular diatreme that consists of a hypabyssal plug and a few dikes exposed among blocks of shale and talus. The Geyser Spring locality consists of two lamprophyre dikes, one 260 feet (80 m) long and the other 690 feet (210 m) long. The exposures may represent segments of a single dike: both dikes trend between N75°E and N85°E, dip steeply, and range in thickness from 12 to 24 inches (30 to 60 cm). The dikes exhibit minor offsets, short branches, and small "blows" that are typical of lamprophyre and kimberlite dikes (see Rock, 1991). The Johnson Ranch intrusion consists of a 2-mile (3.2 km)-long, *en echelon* system of steeply dipping dikes 20 to 40 inches (50 to 100 cm) thick that trend N70°E. A subcircular diatreme 200 feet (60 m) in diameter containing bedded tuffs and country rock xenoliths also is preserved at Johnson Ranch. Gold Butte is an erosional remnant of a diatreme that consists of a small hypabyssal plug and dikes set in a matrix of sedimentary rocks, talus, and minor tuffs. Thin dikes, generally < 20 inches (50 cm) thick, extend intermittently east and west for 1.5 to 2.0 miles (2 to 3 km) from the butte. These have a dip near vertical, a strike from N80°E to East and, with few exceptions, show no topographic relief. Blows, splays, and horned terminations are common features among the dikes. Igneous rocks exposed in Porcupine Dome are classified either as crystalline varieties (plugs, most dikes, and rare sills) or as fragmental types (diatremes [pipes] and local blows in dikes). Further details are given in Doden and Gold (1993b).

The best preserved igneous rocks occur in small outcrops at Froze-to-Death Butte and Gold Butte. These rocks contain the smallest amounts of carbonate minerals, typically less than 50% by volume, and are regarded as "least-altered" for convenience. Almost all other igneous rocks in these and other Porcupine Dome intrusions contain much more carbonate (> 50%) and fewer primary silicate minerals. These carbonate-rich rocks comprise dikes and breccia matrices. Least-altered rocks are readily distinguished from their more altered counterparts: the former are dark gray, hard, and appear fresh in hand specimen, whereas the most-altered rocks are typically buff-orange, friable, and contain few visible silicate minerals.

Sampling and analytical procedures

Hand specimens of crystalline and fragmental rocks were collected from outcrops at all four Porcupine Dome intrusions for petrographic study. Crystalline rocks containing minimal foreign rock contaminants were chosen for whole-rock chemical analysis. Details concerning the analytical procedures are discussed in

Doden (1996). In previous petrological studies (Hearn, 1989b; Scambos, 1991) whole-rock samples were collected from the best preserved outcrops at Froze-to-Death Butte and Gold Butte diatremes. No kimberlite indicator minerals were previously reported from any Porcupine Dome intrusion. In this work a few skeletal garnets were observed in one thin section made from the freshest rocks available at Froze-to-Death Butte. Garnets with kelyphitic alteration rims also were found within an outcrop of more altered rock at the same intrusion. The greatest concentration of garnets, however, was recovered from ant hills on Froze-to-Death Butte. Harvester ants collect granules from 0.04 to 0.20 inches (1 to 5 mm) in diameter to construct the outside of their hills, which are low mounds typically 20 to 40 inches (50 to 100 cm) across. Garnets were also recovered from ant hills on the Johnson Ranch diatreme. Picroilmenites (magnesian ilmenites) were collected from the same ant hills that contained the garnets at both Froze-to-Death Butte and Johnson Ranch. Kimberlite indicator minerals were not present in ant hills examined at Gold Butte or Geyser Spring, although other workers have recovered pyrope garnets from Gold Butte (W.D. Hausel, Wyoming State Geological Survey, personal communication, 1996). Other common kimberlite indicator minerals, such as Cr-diopside, were not observed in outcrop or in any of the ant hills investigated in this study. In addition, no co-existing peridotitic mineral pairs such as garnet-pyroxene or garnet-spinel were found.

Garnets and picroilmenites were readily identified in the field and approximately 5 to 10 pounds (3 to 5 kg) of granules were scooped from each ant hill sampled. The minerals were separated by washing the ant hill debris, then sorting by hand using a binocular microscope to discriminate grains by color, transparency, and luster. Approximately 85 garnet grains or fragments of grains of all compositions were recovered from Froze-to-Death Butte and about 30 were collected from Johnson Ranch. Purple garnets comprise a small subset (six) and their numbers may not afford the best statistical representation of the true garnet xenocryst population. Consequently, garnets of all colors and sizes were analyzed. Furthermore, the garnets sampled in this study are a subset of those "sampled" by the ants; they are not necessarily representative of all Porcupine Dome garnets.

Garnets and ilmenites were mounted in epoxy, polished, and examined under reflected light prior to analysis with the electron microprobe. The grains were analyzed using a Camebax (Cameca) Microbeam™ fully automated electronprobe microanalyzer at Pennsylvania State University (PSU). Typical operating conditions included an accelerating voltage of 15 kV, a beam current of 12 nanoAngstrom units (nÅ), and a beam spot size of 2 micrometers (μm). A range of natural and synthetic standards from the PSU Mineral Constitution Laboratory's collection was used for standardization. A Cameca PAP version of standard ZAF matrix corrections was applied to the data. Stoichiometry was assumed for all mineral phases and ferric iron contents were calculated on-line using the method of Carmichael (1967). At least two analyses per grain were made for both garnets and ilmenites. At least four analyses were made on randomly selected grains to test for compositional zoning.

Results

Important aspects of the mineralogy and chemistry of Porcupine Dome igneous rocks and indicator minerals are discussed below. A brief summary of petrography is provided by Doden and Gold (1993b). The following discussion includes some aspects of mineral compositions not previously available. Further details concerning petrography, mineralogy, and chemistry are given in Doden (1996).

Petrography and mineralogy

Least-altered igneous rocks of Froze-to-Death Butte and Gold Butte have only minor mineralogical differences and otherwise resemble one another compositionally and texturally. Phlogopite and olivine phenocrysts and xenocrysts are set in a matrix of phlogopite, calcite, dolomite, spinels, serpentine, chlorite, apatite, and ± perovskite. Minor amounts of pyrope and picroilmenite xenocrysts occur within the rocks of Froze-to-Death Butte. Inequigranular textures of these rocks are characteristic of many kimberlites (Mitchell, 1986) and ultramafic lamprophyres (Rock, 1987). Mitchell and Bergman's (1991) classification of Porcupine Dome rocks as "olivine-phlogopite-diopside lamproites" implies that diopside is present, but none was observed in this study or by Scambos (1991) or Hearn (1989b).

Olivines in the Porcupine Dome rocks have a bimodal size distribution and are classified as macro- or microcrysts: the former are ≥ 0.2 inches (5 mm) in size and the latter are ≤ 0.08 inches (2 mm). Macrocrysts are visible in some hand specimens and were ob-

served in one thin section from Froze-to-Death Butte. Olivine in this group ranges in size from 0.2 to 0.5 inches (5 to 12 mm) and has anhedral, subspherical to elongated shapes, and exhibits fractures characteristic of olivine. All olivine crystals have been replaced entirely by calcite, serpentine, and bowlingsite (?). The modal abundance of olivine in this size fraction is low; volumetrically the crystals comprise less than 5% of the Froze-to-Death rocks and none have been confidently identified in other Porcupine Dome intrusions. No olivine macrocrysts are preserved adequately for microprobe analysis. The abundant microphenocrysts [0.008 to 0.03 inches (0.2 to 0.7 mm)] of olivine are regarded as microcrysts, and are identified in thin section primarily on the basis of crystal morphology, fractures, and typical replacement products. Most crystals of olivine are completely replaced by serpentine and chlorite; others are replaced by combinations of calcite, dolomite, phlogopite, and quartz. Partially preserved olivine microcrysts in Gold Butte range in composition from Fo_{83} to Fo_{89}.

Two textural varieties of phlogopite (Mg/Fe ratio > 2) are evident in least-altered rocks: phenocrysts/xenocrysts and matrix phlogopite. The former consist of relatively large crystals that range in size from 0.01 to 0.08 inches (0.25 to 2.0 mm). These crystals commonly are kinked; most are anhedral and have well-rounded margins. Although the large phlogopites appear to be phenocrysts, the crystal morphologies and the complex and mixed chemical zoning trends identified from electron microprobe data confirm that some are xenocrysts. The smaller [< 0.01 inch (0.25 mm)] phlogopite crystals are tabular to lath-shaped microphenocrysts that form part of the groundmass. None are kinked and most are subhedral to euhedral.

The compositions of Porcupine Dome phlogopites are typical of those found in many kimberlites and ultramafic lamprophyres (**Figure 2a**). These micas have moderately high concentrations of Al_2O_3 (11 to 17 wt. %), although concentrations of TiO_2 (many crystals > 3 wt. %) reach higher levels than in most kimberlites. Kimberlite phlogopites are generally zoned (core-to-rim) toward decreasing TiO_2 but with variable Al_2O_3, whereas those in Porcupine Dome rocks are zoned toward increasing TiO_2 and Al_2O_3. In general, these compositions and zoning trends better match those of phlogopites in ultramafic lamprophyres. Lamproite phlogopites (e.g., Smoky Butte, Prairie Creek) have distinct compositions, generally lower Al_2O_3 and higher TiO_2 than Porcupine Dome phlogopites (**Figure 2a**), with core-to-rim crystal zoning of low to high TiO_2 and high to low Al_2O_3 (Mitchell and Bergman, 1991).

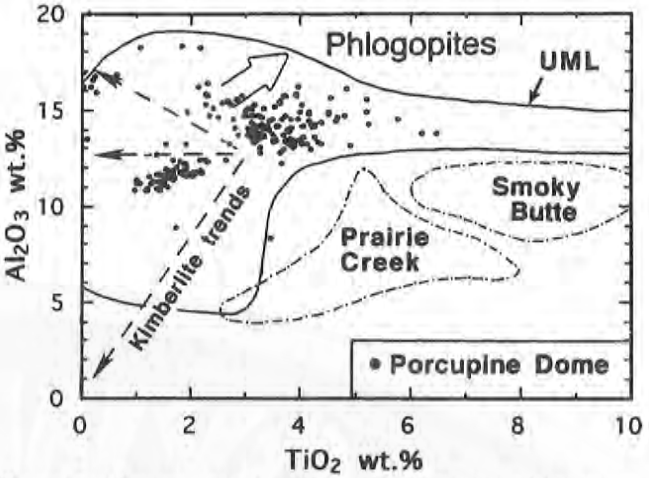


Figure 2a. Concentrations of Al_2O_3 versus TiO_2 in phlogopites from Porcupine Dome intrusions, east-central Montana. Broad arrow indicates dominant direction of compositional zoning (core-to-rim) in Porcupine Dome crystals. Shown for comparison are the range in phlogopite compositions from ultramafic lamprophyres (UML) worldwide (Rock, 1991), the range and zoning trends for kimberlite phlogopites (Mitchell, 1986), and the range in compositions for phlogopites from Smoky Butte and Prairie Creek lamproites (Mitchell and Bergman, 1991).

Opaque minerals in the least-altered rocks of Porcupine Dome consist mostly of magnetite and titanian magnetite; some of these crystals have chromite or aluminous magnesian chromite cores that are rimmed by titanian magnetite. Many of these crystals are partially or completely altered to hematite and/or leucoxene. Limonite occurs locally, and is especially evident in more carbonate-rich rocks as irregularly-shaped stains. Porcupine Dome spinels compositionally resemble those of ultramafic lamprophyres, but share some characteristics with kimberlite spinels and, to a lesser extent, with those of olivine lamproites. Porcupine Dome spinels have Fe/(Fe+Mg) ratios (where Fe=total iron as Fe^{2+}) that partially overlap the "magmatic Trend-2" compositional field of kimberlite spinels (Mitchell, 1986), with some transition toward "Trend-1" kimberlite compositions, a chemical characteristic common among spinels in ultramafic lamprophyres (Rock, 1991). However, Porcupine Dome spinels generally have < 1 wt. % MnO, similar to those in kimberlites, rather than 1 to 2.5 wt. % MnO that is typical of lamprophyre spinels (Mitchell, 1979). Values of Cr/(Cr+Al) in Porcupine Dome spinels are mostly < 0.3, much less than the very high ratios [Cr/(Cr+Al) > 0.95] characteristic of lamproite spinels (Mitchell and Bergman, 1991).

Carbonate comprises approximately 30 to 50% of the groundmass and less than 50% total of the least-altered rocks. Calcite, dolomite, and minor ferruginous calcite are present in the matrix of least-altered rocks,

with calcite typically dominating over dolomite by as much as 4:1. In addition to other textures, calcite occurs in the matrix as mosaics intergrown with euhedral phlogopites and spinels. Such intergrowths and the relatively high concentrations of Sr (0.10 to 0.80 wt. % SrO) in some matrix calcite may indicate a primary magmatic origin (e.g., Brookins and Watson, 1969; Scott Smith and Skinner, 1984; Hamilton and Rock, 1990; Bourne and Boss, 1991; Pope and Thirlwall, 1992) for the groundmass carbonate. Globular, calcite-filled structures up to 0.59 inch (15 mm) across resemble nodular masses of calcite found in some kimberlites and may be the fillings of amygdules (see Kirkley and others, 1989) or replacements of olivine macrocrysts.

Whole-rock chemistry

Whole-rock chemical analyses of Porcupine Dome igneous rocks are presented in Doden and Gold (1993b) and Doden (1996). The least-altered rocks are ultramafic ($SiO_2 \leq 25$ wt. %; MgO generally > 15 wt. %) and have low to moderate concentrations of Al_2O_3 (2.7 to 3.2 wt. %), K_2O (1.7 to 1.8 wt. %), and Na_2O (0.06 to 0.22 wt. %) compared to most alkaline rocks. Ratios of K_2O/Na_2O are anomalously high in most samples analyzed (all > 3.0), but this is due primarily to the low Na_2O. By the criteria of Foley and others (1987), Porcupine Dome igneous rocks would be considered ultrapotassic, but because their K_2O concentrations are less than 3.0 wt. %, they are regarded as potassic. Least-altered rocks of Porcupine Dome also are peraluminous (molar $K_2O/Al_2O_3 < 1.0$). Whole-rock major element concentrations overlap those of ultramafic lamprophyres and African kimberlites. Porcupine Dome compositions resemble those of kimberlites, alnoites, and peridotites of the Missouri River Breaks field but are different from those of lamproites (**Figure 2b**). Lamproites of Smoky Butte and the Leucite Hills (**Figure 1**) have higher concentrations of SiO_2 (commonly > 50 wt. %), TiO_2 (4 to 6 wt. %), Al_2O_3 (8 to 10 wt. %), and K_2O (6 to 12 wt. %) (Mitchell and Bergman, 1991) than the Porcupine Dome rocks.

The least-altered rocks of Porcupine Dome are rich in incompatible trace elements, notably Ba, Sr, and the light rare earth elements. Concentrations best resemble those in kimberlites and ultramafic lamprophyres, but are significantly less than those of most lamproites. Barium (up to 7200 ppm) and Sr (up to 2383 ppm) levels, however, are unusually high for ultramafic lamprophyres (see Rock, 1991). Incompatible element spiderdiagram patterns for Porcupine Dome rocks are remarkably similar to patterns for

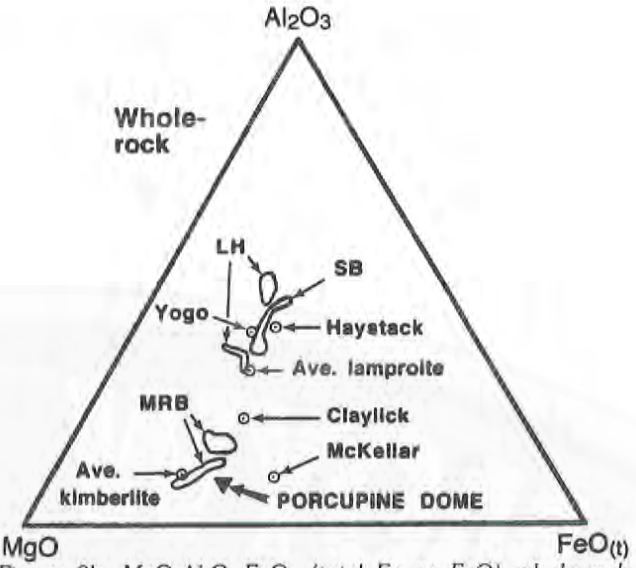


Figure 2b. MgO-Al_2O_3-FeO_T (total Fe as FeO) whole-rock compositions for least-altered igneous rocks of Porcupine Dome (inverted triangle) and elsewhere. Average kimberlite composition from Dawson (1967) and average lamproite composition from Bergman (1987). Abbreviations: Claylick, ultrabasic lamprophyre from Claylick Creek, Kentucky (Koenig, 1956); Haystack, monticellite peridotite from Haystack Butte, Montana (Hearn, 1989b); LH, Leucite Hills lamproites (Mitchell and Bergman, 1991); McKellar, ultrabasic lamprophyre from McKellar Harbor, Ontario (Platt and Mitchell, 1982); MRB, kimberlites (lower field) and alnoites and peridotites (upper field) from the Missouri River Breaks, Montana (Hearn, 1989b); SB, lamproites from Smoky Butte, Montana (Mitchell and Bergman, 1991); and Yogo, ouachitite lamprophyre from the Yogo dike, Little Belt Mountains, Montana (Meyer and Mitchell, 1988).

kimberlites and carbonate-rich mica peridotites of the Missouri River Breaks (Hearn, 1989b), although the former show minor enrichments in Ti, Th, Nb, and La relative to Missouri River Breaks kimberlites.

Indicator minerals

Garnets

Garnets from Froze-to-Death Butte and Johnson Ranch were classified according to color as observed under a binocular microscope. The garnets comprise a rather narrow size range, from 0.04 to 0.2 inches (1 to 5 mm) in maximum dimension, as do most of the geological and biological materials that comprise ant hills. Although some grains exhibit rough, pitted surfaces that commonly have kelyphitic alteration rinds, most garnet grains show fresh conchoidal fracture surfaces, suggesting they are fragments of larger crystals. Garnet grains (or fragments) fall into one of two color categories, herein designated as **red-purple** and **pink-orange** according to the dominant colors observed. If all garnet fragments are considered, including a group of ~0.04 inches (~1 mm) slivers (< 10% of the total

population recovered), the colors present in all sizes of garnets range almost continuously from deep purple to pink to nearly colorless. However, the lightest pink and orange garnets are also the smallest fragments, commonly thin slivers, and it is believed that in thicker grains they would be darker. This hypothesis is supported by the compositional data that shows that some light pink fragments are chemically similar to those of the red-purple group. Macroscopically, no differences are evident between garnets from Froze-to-Death Butte and Johnson Ranch.

Garnets from Froze-to-Death Butte and Johnson Ranch are considered together in the following discussion as no significant compositional differences exist between crystals from the two intrusions. Electron microprobe analyses demonstrate that two distinct compositional groups exist. The red-purple garnets are significantly enriched in Cr_2O_3, MgO, and CaO relative to the pink-orange garnets (**Table 1**). The bimodal distribution of garnet compositions is shown in **Figures 3a** and **3b** in terms of mole % Mg, Ca, and Fe (recalculated to 100%). Compositions of Porcupine Dome red-purple garnets (Mg-rich group) fall within a relatively restricted range that overlaps that of garnets from the Lesotho kimberlites (**Figure 3a**) and those from the Williams kimberlite of the Missouri River Breaks diatreme field in north-central Montana (**Figure 3b**). However, the Porcupine Dome garnets proportionally are not as Mg-rich as garnets found as some inclusions in diamonds (e.g., Meyer and Boyd, 1972). Red-purple garnets of Porcupine Dome also overlap compositionally with Cr-rich garnet megacrysts from the Sloan-Nix kimberlites of the State Line district (Eggler and others, 1979) and may correspond to the "Cr-rich megacryst trend" (see Mitchell, 1986). Although no mantle-derived xenoliths have been found within the Porcupine Dome diatremes, compositions of the red-purple garnets are similar to those derived from lherzolites. Some red-purple garnets are enriched in the knorringite component [(Cr/Cr+Al) x 100 = 16 to 20], compositions that may be indicative of formation under high pressures (Irifune and others, 1982), perhaps approaching 45 to 50 kb.

Porcupine Dome garnets are classified into "cluster groups" using the statistical scheme developed by Dawson and Stephens (1975) based on their studies of South African garnets. Although not without its limitations (see Mitchell, 1986), the Dawson and Stephens (1975) classification scheme continues to be employed (e.g., McGee, 1988; Simpson, 1993) and provides a general basis for chemical comparison with garnets from other localities. All average oxide values for Porcupine Dome garnets lie within the acceptable range for the Dawson and Stephens system, with the

Table 1. Chemical composition of garnets and picroilmenites from Froze-to-Death Butte, Porcupine Dome.

	Garnets				Picroilmenites		
	Red-purple		Pink-orange				
	Mean[1]	Range	Mean[2]	Range	Mean[3]	Typical core	Typical rim
SiO_2	42.48	41.85-43.30	38.58	38.05-39.38	-	-	-
TiO_2	0.23	0.20-0.26	0.02	0-0.05	47.61	48.24	49.52
Fe_2O_3	0.91	0.51-1.52	0.07	0-0.42	13.57	12.63	14.08
Al_2O_3	18.65	18.16-19.40	21.81	21.22-22.19	0.07	0.07	0.12
Cr_2O_3	6.52	5.53-7.01	0.03	0-0.08	0.21	0.24	0.32
MgO	19.37	19.01-19.74	6.37	5.90-6.51	6.59	7.25	11.86
CaO	6.45	6.00-6.75	2.08	1.72-2.18	-	-	-
MnO	0.30	0.22-0.34	1.31	1.21-1.41	0.23	0.20	0.45
FeO	6.49	5.96-6.93	30.55	30.24-30.86	30.99	30.41	23.17
Na_2O	0.03	0.02-0.05	0.01	0-0.03	-	-	-
Total	101.44	100.55-102.03	100.83	100.12-101.67	99.27	99.05	99.53

[1] Mean of analyses from 9 crystals

[2] Mean of analyses from 11 crystals

[3] Mean of analyses from 7 crystals

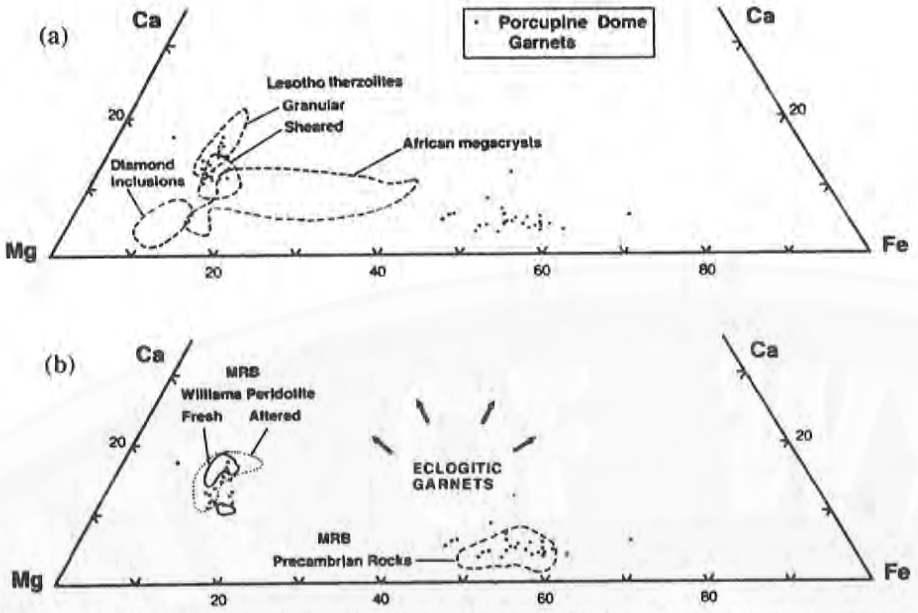


Figure 3. Compositions of garnets in terms of Mg-Ca-Fe cations. Points represent averages of three or more analyses for individual garnet grains from Froze-to-Death Butte and Johnson Ranch diatremes. (a) Porcupine Dome and African garnets. Red-purple Porcupine Dome garnets lie within the Mg-rich cluster of points and pink-orange garnets comprise the more Fe-rich cluster. Compositional fields for various African garnet sources (data from Reid and Hanor, 1970; Meyer and Boyd, 1972; Boyd and Nixon, 1972). (b) Porcupine Dome garnets compared to those from the Missouri River Breaks (MRB) diatremes of north-central Montana (data from Hearn and McGee, 1983). MRB Precambrian Rocks field indicates range of compositions in garnets derived mainly from schists.

exception of FeO_T (total Fe as FeO) for Johnson Ranch pink-orange crystals. Using average Froze-to-Death Butte and Johnson Ranch oxide values, the red-purple garnets are classified with Dawson and Stephens' (1975) flow sheet as part of the G9 cluster group (chrome pyrope), typical of garnets from garnet lherzolite nodules found in kimberlites. Red-purple garnets from both intrusions resemble G9 garnets in terms of mean TiO_2 (0.12 wt. %), MgO (19.36 to 20.13 wt. %), and CaO (5.60 to 5.72 wt. %), but more closely resemble G10 (sub-calcic chrome pyrope) garnets in their concentrations of Cr_2O_3 (5.62 to 5.95 wt. %). Red-purple Porcupine Dome garnets compositionally overlap garnets from South African kimberlite nodules (**Figure 3a**) as well as peridotite nodules from the Missouri River Breaks (**Figure 3b**).

Porcupine Dome garnet CaO and Cr_2O_3 compositions are shown in **Figure 4**. The vertical dashed line separates eclogitic garnets (< 2.0 wt. % Cr_2O_3) from peridotitic garnets (> 2.0 wt. % Cr_2O_3) as proposed by Gurney (1984). Red-purple Porucpine Dome garnets fall within the high-Cr peridotitic field (G9 compositions; Gurney, 1984) and overlap with compositions of the Williams kimberlite garnets (Missouri River Breaks), although the latter extend to somewhat higher Cr_2O_3 contents. The trend observed in **Figure 4** corresponds to that found in garnet lherzolites (e.g., Sobolev and others, 1973; Gurney, 1984). Although most Porcupine Dome red-purple garnets are G9 in composition, two crystals from Froze-to-Death Butte lie very near the G9-G10 boundary. One crystal from Johnson Ranch clearly lies within the G10 field and is subcalcic (harzburgitic). Subcalcic chromian garnets are regarded as evidence for a diamond-bearing harzburgitic source (e.g., Sobolev, 1977, 1984; Gurney, 1984; McCallum and Waldman, 1991).

Pink-orange garnets from Porcupine Dome have considerably more Fe and less Mg than the red-purple garnets (**Table 1**). The somewhat broader range of pink-orange Porcupine Dome garnets closely matches that of similarly-colored garnets recovered by Hearn and McGee (1984) from the Williams kimberlite diatreme. Some garnet compositions from the Williams diatreme, however, are substantially more Ca- and Fe-rich than those of Porcupine Dome, a factor attributed to either garnet xenocrysts from other sources (as suggested by Hearn and McGee, 1983), or the fact that a considerably larger garnet population was sampled in the Missouri River Breaks diatremes. In other words, it is possible that the Porcupine Dome garnet population sampled is not entirely representative of all types of garnet brought up by the diatremes.

Classification of the pink-orange garnets by Dawson and Stephen's (1975) flow sheet procedure shows that the mean Froze-to-Death compositions fall into the G3 cluster group (calcic pyrope almandine) and those of Johnson Ranch in the G5 group (magnesian almandine). By this method the pink-orange compositions correspond to garnets derived from eclogites (Dawson and Stephens, 1975). Although eclogitic garnets span a wide range of compositions, they generally have higher proportions of Ca than garnets from crustal sources. Schultze (1993) concluded that eclogitic garnets typically have between 3.5 and 20 wt. % CaO, and those with less than 3.5 wt. % CaO are probably derived from pelitic crustal rocks. As shown

in **Figure 4**, most Porcupine Dome pink-orange garnets (Cr-poor compositions) have low concentrations of CaO and, by Schultze's (1993) criterion, resemble crustal garnets. Also, in terms of atomic Mg-Ca-Fe, these garnets overlap similarly-colored Missouri River Breaks garnets derived from Precambrian crustal rocks (mainly schists; Hearn and McGee, 1983) as illustrated in **Figure 3b**.

Compositional ranges for North American and worldwide eclogitic garnets from Gurney and Moore (1993) are shown in a plot of Na_2O versus TiO_2 (**Figure 5**). Eclogitic garnets associated with diamonds typically have elevated concentrations of both elemental oxides, and garnets with $Na_2O \geq 0.07$ wt. % are considered significant in diamond exploration (Gurney and Moore, 1993). With the exception of a few analyses, Porcupine Dome garnets, including the pink-orange variety, have $Na_2O \leq 0.07$ wt. % and $TiO_2 \leq 0.4$ wt. %. If any of the pink-orange garnets are eclogitic, then this comparison suggests that they were not derived from a diamondiferous source.

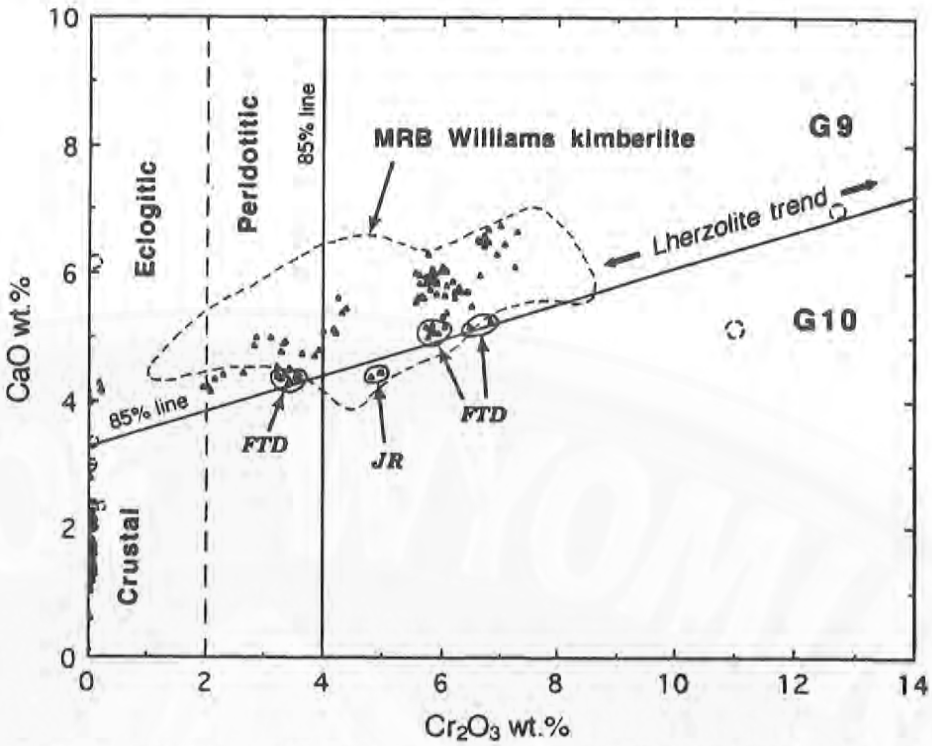


Figure 4. Concentrations of CaO versus Cr_2O_3 for Porcupine Dome (solid triangles for single microprobe analyses) and other garnets including those from Froze-to-Death Butte (FTD) and Johnson Ranch (JR). The solid diagonal and vertical lines separating calcium-rich G9 garnets from sub-calcic G10 garnets were defined by Gurney (1984) on the basis that 85 % of peridotitic garnets associated with diamonds plot in the G10 field. The lherzolite trend indicates the range in compositions typical of garnets from lherzolites (Sobolev and others, 1973). The vertical dashed line at 2.0 wt. % Cr_2O_3 is Gurney and Moore's (1993) division between eclogitic and peridotitic garnets. Most pink-orange garnets of Porcupine Dome lie below the G9-G10 boundary at low Cr_2O_3 levels and correspond to garnets derived from crustal sources (Schultze, 1993), whereas a few crystals of > 3.5 wt. % CaO may be eclogitic. The dashed boundaries show the range in composition of garnets from the Williams kimberlite in the Missouri River Breaks (MRB) diatreme field of north-central Montana (McGee, 1988). Note the small MRB data fields at very low and high Cr_2O_3 values. Most Porcupine Dome peridotitic garnet compositions lie within the lherzolite trend; a few are "G10" in composition or are borderline on the basis of Gurney's (1984) 85% dividing line.

Picroilmenites

Picroilmenite crystals comprise the second type of kimberlite indicator mineral discovered in the Froze-to-Death Butte and Johnson Ranch diatremes. Picroilmenites are at least ten times more abundant than garnets in ant hills from both diatremes. Most crystals range in size from 1 to 5 mm and have external surfaces that are heavily pitted due to abrasion and/or alteration to leucoxene. Others have the characteristic conchoidal fracture and metallic luster. Picroilmenites have not been observed in thin section from any crystalline textured igneous rock from Porcupine Dome.

Ranges in the composition of representative picroilmenite macrocrysts from Porcupine Dome intrusions are summarized in **Table 1**. In terms of mole % Mg-Mn-Fe, the picroilmenites overlap those of ultramafic lamprophyres and kimberlites (**Figure 6a**), but a plot of mole % Fe^{3+}-Fe^{2+}-Mg (**Figure 6b**) suggests that the closest affinity is with kimberlite ilmenites. Microprobe traverses across ilmenite xenocrysts show that MgO concentrations are constant throughout the grains, but are enriched significantly at the margins (up to 5 wt. %), typical of picroilmenite xenocrysts in kimberlites (Mitchell, 1986).

In contrast to garnets, the rarity of ilmenite inclusions in diamonds provides an insufficient data base to define any useful discriminating compositional coefficients (Haggerty, 1991; Moore and others, 1991). Nevertheless, the variations in MgO versus Cr_2O_3 in

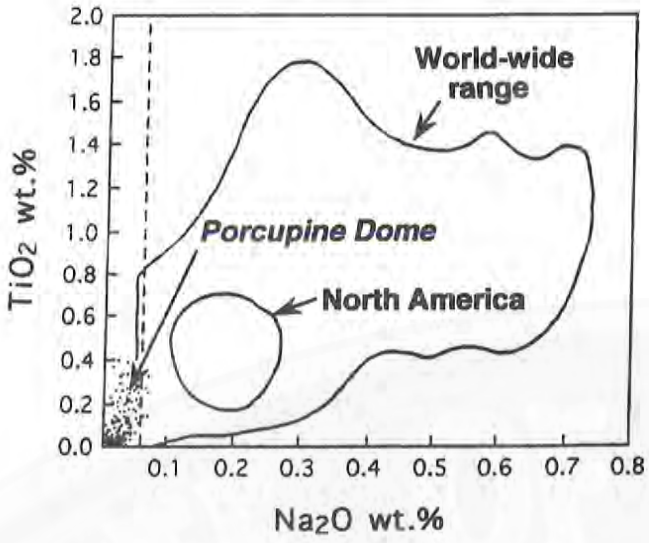


Figure 5. Concentrations of Na_2O versus TiO_2 for Porcupine Dome garnets (points include both red-purple and pink-orange crystals) and in eclogitic garnets elsewhere. With the exception of a few analyses, all Porcupine Dome garnets have $Na_2O \leq 0.07$ wt. % and $TiO_2 \leq 0.4$ wt. %. Compositional ranges for North American and world-wide eclogitic garnets from Gurney and Moore (1993).

ilmenites have been used to evaluate the potential presence of diamonds (e.g., Haggerty, 1991; McCallum and Vos, 1993). Porcupine Dome ilmenite compositions are plotted in **Figure 7**, along with ilmenite compositions from kimberlites. Concentrations of Mg and Cr in picroilmenites vary considerably among and within kimberlite districts (Mitchell, 1986) but the compositional trends commonly form broad parabolic patterns in plots of MgO vs. Cr_2O_3 (Haggerty, 1975). According to Haggerty (1991), the presence of diamond in kimberlites can be correlated with high concentrations of MgO and Cr_2O_3 in ilmenite macrocrysts of the same rocks. Kimberlitic ilmenites with high MgO and Cr_2O_3 (e.g., Sloan-Nix, **Figure 7**) imply lower Fe^{3+}/Fe^{2+} ratios and hence relatively reducing environments conducive to diamond survival. Ilmenites with low MgO concentrations (e.g., Iron Mountain) imply higher Fe^{3+}/Fe^{2+} ratios and oxidizing conditions, unsuitable for diamond preservation (McCallum and Waldman, 1991). Porcupine Dome ilmenites have only moderate MgO concentrations and particularly low amounts of Cr_2O_3 relative to kimberlitic ilmenites (**Figure 7**).

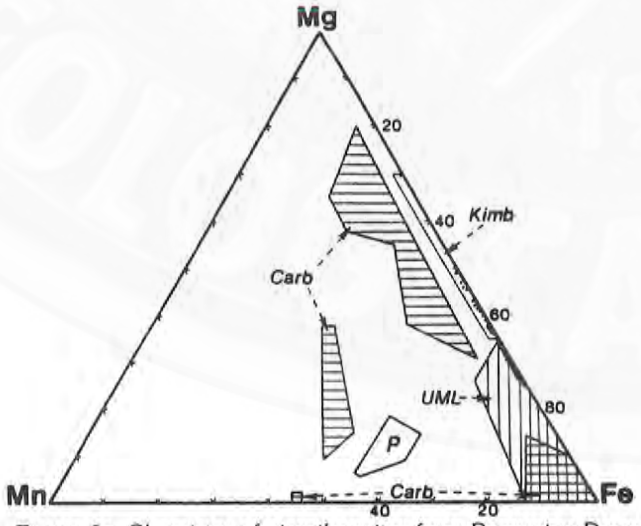


Figure 6a. Chemistry of picroilmenites from Porcupine Dome in mole % Mg-Mn-Fe (data points along Mg-Fe axis). Compositional field for ultramafic lamprophyre (UML) ilmenites from Rock (1986). Fields for kimberlite (Kimb), carbonatite (Carb), and Premier calcite kimberlite (P) ilmenites from Mitchell (1986) and references therein. After Hamilton and Rock (1990).

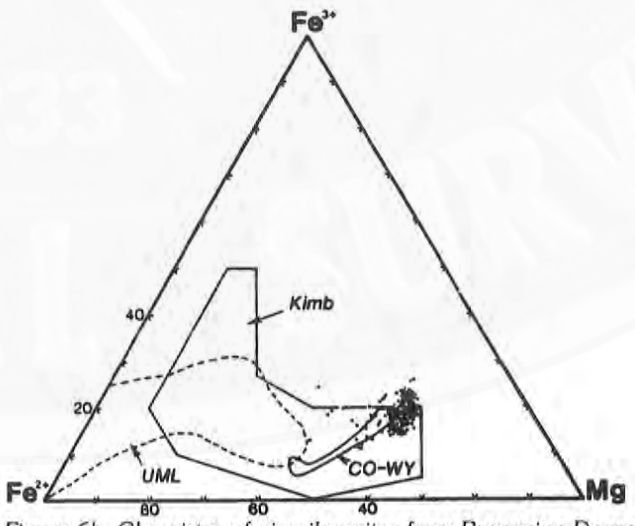


Figure 6b. Chemistry of picroilmenites from Porcupine Dome in mole % Fe^{3+}-Fe^{2+}-Mg. Compositional fields for ultramafic lamprophyre (UML) ilmenites from Rock (1991), kimberlite (Kimb) ilmenites from Mitchell (1986), and Colorado-Wyoming (CO-WY) kimberlite ilmenites from McCallum and others (1975).

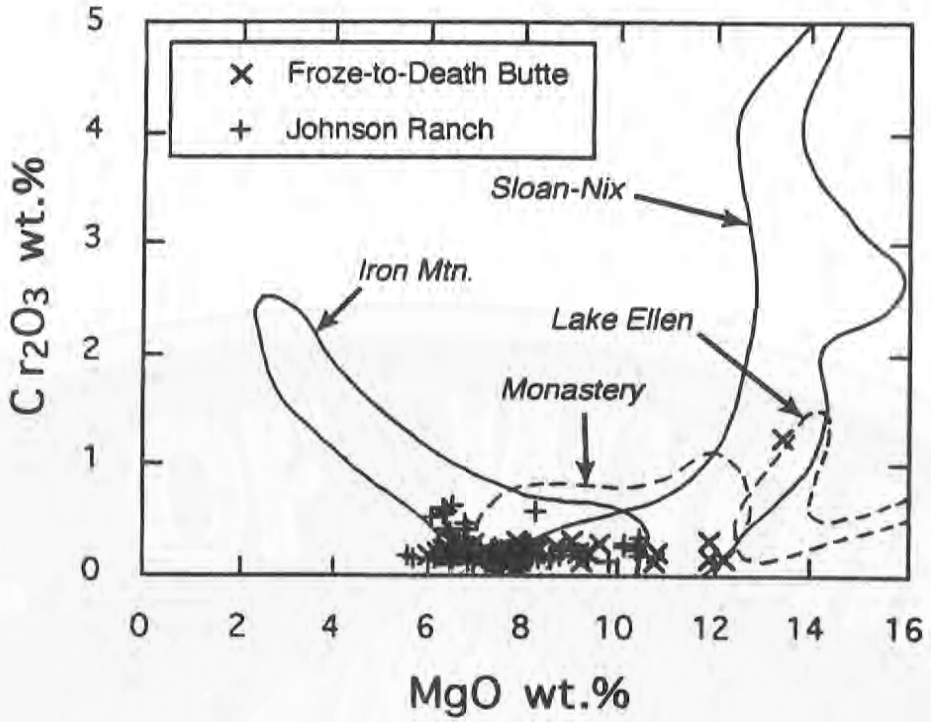


Figure 7. Concentrations of MgO versus Cr_2O_3 in ilmenite xenocrysts from Porcupine Dome. Compositional ranges for megacrysts from some United States (Lake Ellen, Iron Mountain, and Sloan-Nix) and African (Monastery) kimberlites are indicated for comparison (data from McCallum and Vos, 1993). Kimberlitic ilmenites with high MgO and Cr_2O_3 (e.g., Sloan-Nix) imply lower Fe^{3+}/Fe^{2+} ratios and hence relatively reducing environments conducive to diamond survival. Ilmenites with low MgO concentrations (Iron Mountain) imply higher Fe^{3+}/Fe^{2+} ratios and oxidizing conditions unsuitable for diamond preservation (McCallum and Waldman, 1991).

Discussion and conclusions

Rock classification

Previous studies have classified the igneous rocks of Porcupine Dome as carbonate-rich mica peridotites (Hearn, 1989b; Scambos, 1991) or as olivine-phlogopite-diopside lamproites (Mitchell and Bergman, 1991). The Porcupine Dome intrusions have some petrographic and chemical characteristics of kimberlites. The presence of two textural varieties of olivine in Froze-to-Death Butte rocks, for example, is particularly suggestive. However, compared to typical kimberlites (e.g., Mitchell, 1986), olivine macrocrysts are rare, and the presence of only one textural type of olivine suggests a non-kimberlite origin (Clement and others, 1984). In detail, these rocks bear little resemblence to lamproites. Their whole-rock compositions are clearly outside the range of lamproites (see Bergman, 1987), with the exception of some olivine lamproites (e.g., Prairie Creek, Arkansas; Scott Smith and Skinner, 1984). Furthermore, unusual minerals peculiar to lamproites (e.g., priderite, magnophorite, wadeite) do not occur in Porcupine Dome rocks. Also absent are primary quartz, feldspars, amphiboles, pyroxenes, nepheline and other feldspathoids, and glass. The carbonate minerals so abundant in central Montana rocks, many kimberlites, and some ultramafic lamprophyres, are rarely found in lamproites (Bergman, 1987; Mitchell and Bergman, 1991). The geological and petrographic characteristics of the Porcupine Dome igneous rocks are typical of ultramafic lamprophyres. Furthermore, comparisons of mineral zoning trends and whole-rock compositions plotted on major-element oxide discrimination diagrams (Rock, 1991) indicate that these rocks are best classified as **aillikites**, an uncommon variety of ultramafic lamprophyre (Rock, 1987, 1991). Alternatively, Mitchell (1994) prefers the name "olivine-phlogopite-(or phlogopite-olivine)-calcite melnoite."

Diamond potential of the Porcupine Dome intrusions

Ultramafic lamprophyres should not be disregarded as potential sources of diamond, as microdiamonds

have been found in ultramafic intrusions near Kimberley, South Africa (Clement and others, 1979) and in the Maude Creek lamprophyres of Western Australia (Jaques and others, 1986; Lucas and others, 1989). Other ultramafic lamprophyres may contain diamonds (Rock, 1991). Although these deposits are not economical, the resemblance of many ultramafic lamprophyres to kimberlites with regard to tectonic setting and compositions suggests that their magmas also had the potential to sample diamondiferous mantle. Moreover, certain diamond-bearing intrusions previously identified as kimberlites may in fact be varieties of ultramafic lamprophyres (Rock, 1991).

None of the intrusions from the Missouri River Breaks, Grassrange, or Porcupine Dome have been demonstrated to be a source of diamonds (McGee, 1988; Doden, 1996), but a few alluvial diamonds whose sources are yet unknown have been recovered elsewhere in Montana (Hausel, 1996). Many aspects of the Porcupine Dome intrusions are considered favorable indicators for diamond, including their geological setting. The northern Wyoming craton has an age (>2.5 Ga; e.g., Peterman and Futa, 1987), thickness [>95 miles (>150 km); Eggler and Furlong, 1991], and tectonic history (Green and others, 1985; Houston and others, 1993) typical of others that host kimberlites and lamproites (e.g., western Liberia craton, Haggerty, 1982; Brazil craton, Svisero and others, 1979). Such thick and ancient cratons are believed to be a prerequisite for the generation of magmas from within the diamond stability field (e.g., Haggerty, 1986). Evidence from garnet peridotite and garnet orthopyroxenite nodules in the Missouri River Breaks intrusions confirms that kimberlite magmas in this field originated from depths exceeding 180 km (Hearn and McGee, 1984). The mineralogical and chemical composition of Porcupine Dome aillikites, and their similarities to Missouri River Breaks kimberlites (Doden and Gold, 1993b; Doden, 1996), suggest that the magmas are derived from a deep source. The central Montana rocks are typical of many ultramafic lamprophyres and kimberlites in that they have abundant carbon (in the form of carbonates, probably primary) and were emplaced rapidly (Doden, 1996), characteristics favorable for diamond exploration targets as they increase the likelihood of preserving metastable diamond (Gold, 1984; Rock, 1991).

Pink-orange garnets of Porcupine Dome were probably derived from a crustal source such as schist. Thus, they afford no evidence for diamond potential as would eclogitic garnets. However, the red-purple garnets fall within the compositional range of those derived from African garnet lherzolite sources. They also are similar to peridotite and megacryst garnets from kimberlites of the Missouri River Breaks diatreme field in north-central Montana. Chromian pyropes are associated with high-pressure source regions (e.g., Sobolev and others, 1973; Meyer and Tsai, 1976). Red-purple garnets, once entrained in Porcupine Dome magmas, have compositions compatible with an upper mantle source; the large proportion of knorringite component [Cr/(Cr+Al) = 0.15-0.20] in some crystals is indicative of high-pressure crystallization (Irifune and others, 1982).

Most of the Cr-rich pyrope garnets of Porcupine Dome best correspond to the G9 cluster group of Dawson and Stephens (1975), although a few compositions lie within the G10 group comprising low-Ca, Cr-rich garnets. Gurney (1984) and Boyd and Gurney (1982) found that, for kimberlites of South Africa, a correlation exists between the presence of subcalcic (G10) garnets and diamonds. Such a correlation may be applicable to North American occurrences as well (McGee, 1988). Daniels and Gurney (1986) and McGee (1988) found that subcalcic (G10) compositions are restricted primarily to garnets of deep purple color. McGee (1988) hand-picked 137 deep purple garnets from the Williams kimberlite at Missouri River Breaks in an effort to maximize the number with subcalcic compositions. Of these, only four grains were shown to be slightly subcalcic, yet McGee concluded that the presence of diamond at Williams is possible solely on this basis. Although the total number of deep purple garnets sampled from Porcupine Dome diatremes (six whole grains out of 85 total) is much smaller than that found by Hearn and McGee (1983) and McGee (1988), the proportion of subcalcic (G10) to high calcium (G9) purple garnets is considerably greater.

Based on the geology, mineralogy, and chemistry of Porcupine Dome rocks, and the presence of deep purple, subcalcic pyropes (Tompkins, 1987; McGee, 1988), it is concluded that the magmas were derived from a potentially diamondiferous mantle source. However, the apparent oxidation state of the magmas is considered unfavorable evidence for diamond preservation in the Porcupine Dome diatremes. Picroilmenites from Froze-to-Death Butte and Johnson Ranch have intermediate values of MgO and low values of Cr_2O_3 compared to ilmenites typically associated with diamond-bearing rocks (e.g., Gold, 1984; Haggerty, 1991). Were concentrations of MgO higher, lower Fe^{3+}/Fe^{2+} ratios would be implied for the melt and a relatively reducing environment may have existed, one better suited to diamond survival (McCallum and Waldman, 1991). In addition, the igneous rocks have undergone multiple stages of alteration that include the conversion of many spinels to

hematite. This and the relatively high whole-rock Fe_2O_3/FeO ratios in most samples (Doden, 1996) suggest oxidizing conditions. Some of this could be the result of secondary (post-magmatic) effects of groundwater. Even if diamonds survived, the small size of the intrusions presently exposed would severely restrict the available tonnage and, hence, would compromise the economic viability of these deposits.

The richest concentrations of diamond in diamond-bearing kimberlites and lamproites are typically in the upper tuffaceous facies of the pipes (Mitchell and Bergman, 1991). The level of erosion in the Porcupine Dome localities is such that only the hypabyssal facies rocks are preserved, most of the crater facies materials presumably having been removed by erosion, if the standard model of kimberlite diatremes (Hawthorne, 1975) is accepted. The exception may be Johnson Ranch, which has a diatreme containing down-dropped blocks of preserved tuffs. If diamonds were present at one time in Porcupine Dome diatremes, but were restricted to the upper parts of the pipes as in kimberlite and lamproites, then it seems likely that none would be found in the remaining intrusions, because the upper levels have been removed by erosion. It follows that any remaining diamonds would exist only in the surrounding drainage systems. No attempt has been made in this work to investigate nearby stream beds for diamonds or indicator minerals.

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Application of infrared field spectroscopy to surficial mapping[1]

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Abstract

Surficial mapping of industrial minerals can be facilitated using infrared spectroscopy to determine characteristics as they coincide with subsurface conditions. Spectrometer measurements in the field can provide information indicative of potential deposits quickly and accurately.

Based on principles used successfully in chemical analyses, a portable infrared field spectrometer measures the vibrational frequencies of responses inherent to individual materials when exposed to infrared energy. The frequency response is recorded as a spectral curve, which represents the amount of infrared energy reflected at certain wavelengths. Each curve is unique to a given set of molecules because of differences in the reflective properties of a material, and can be used to assess the occurrence of specific industrial or alteration minerals in the field.

Rocks, soils, vegetation, and water are separable within the mid-infrared range (1300 to 2500 nanometers), according to their spectral characteristics. Surficial material can be valuable in this context because its spectral features are often indications of products associated with the weathering and alteration of the underlying materials. The presence of clays and other minerals is particularly identifiable in this range due to the crystalline structures and water retention properties of these minerals. Examples such as clays, zeolites, and minerals found in kimberlitic terrains have been found distinguishable in the mid-infrared spectral range and as such can be mapped effectively with these techniques.

Infrared field spectroscopy can be a worthwhile mapping tool providing extensive field data in a short period of time. When compared to traditional field surveys, spectroscopy is relatively inexpensive and reduces data processing time. This, and the reliability of spectral identification, make spectroscopy a valuable asset to surficial mapping of industrial mineral deposits.

Introduction

Spectroscopy is a technique historically employed as a tool of chemistry. Chemists often utilize the interactions between electromagnetic energy and molecules to accurately identify many compounds of questionable composition. Geologic research has adopted the same principles, but is using recent technology to expand spectroscopy into an advanced field science.

A common utilization of electromagnetic energy is x-ray analysis, a successfully proven laboratory technique. Identification of minerals by XRD (x-ray diffraction) is accomplished by exposing a sample to electromagnetic radiation in the x-ray range of wavelengths (**Figure 1**). The distribution of electrons in the atoms comprising the material is modified (**Figure 2**) according to the chemical and physical properties of

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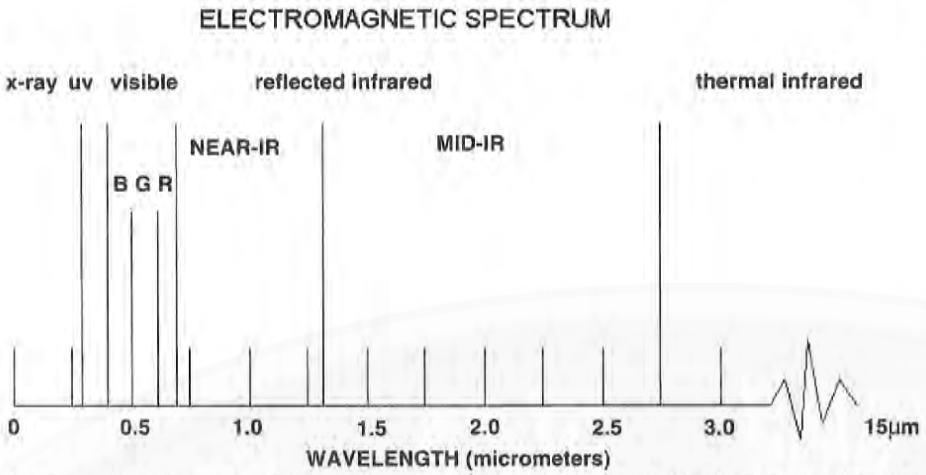


Figure 1. A generalized version of the electromagnetic spectrum. The longer wavelengths in the mid-infrared range have been found to be as useful as the shorter x-rays for mineral identification.

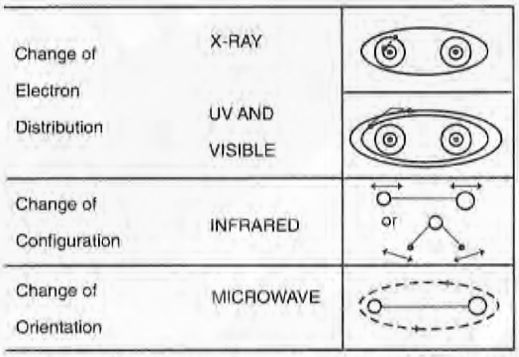


Figure 2. Atomic response to electromagnetic radiation. Note the differences in response to x-rays versus infrared. Infrared energy causes vibrations to occur when bonds bend and stretch. Adapted from Banwell (1983).

the substance. The specific atoms and their configuration define those properties of a mineral and as such can be exploited to identify the mineral on the basis of how its molecules respond to an x-ray.

With the need for scientists to make accurate distinctions between materials in the field, the utility of XRD is limited by its impracticality as a field instrument. Sample identification is significantly hindered by the time required and difficulties encountered in returning the sample to an XRD laboratory for analysis and identification. Also, the samples may be adversely affected by the sampling procedure or by subsequent handling (Moore and Reynolds, 1989). Field researchers have accepted this dilemma as an inconvenient but necessary trade for the reputable results of XRD analysis. Recent technological advances have created instrumentation that preserves the reliability and accuracy of traditional spectroscopy while making it accessible for *in-situ* studies.

By virtue of new technology, spectroscopy has precipitated into fields other than chemistry, emerging especially in geologic remote sensing as a modern tool for discriminating between minerals. Since the 1970s, scientists have applied remote sensing of electromagnetic radiation to the study of surficial materials (Condit, 1970; Hunt and Salisbury, 1970, 1971, 1976; Hunt, 1977, 1979; Richardson and Weigand, 1977; Crowley, 1984, 1986, 1991). As remote sensing gained momentum, spectroscopy was utilized as a form of ground-truth for the interpretation of satellite and other multispectral data sets (Lillesand and Kieffer, 1994). The interest and technology have developed only recently to make field spectroscopy an efficient and versatile method for on-site evaluations of most materials.

The advances that have taken spectroscopy into the field are based on advances in solid-state electronics, continued miniaturization, and an ever-widening spectral range sensed by various instruments. Currently, instruments can record reflected energy in narrow bandwidths over the visible and infrared wavelengths, including the middle infrared region (**Figure 1**). Many minerals of interest are quite definable by their response to energy in this region of the spectrum from 1300 to 2500 nanometers (nm) (Lillesand and Kieffer, 1994; Avery and Berlin, 1992; Sabins, 1983).

Mid-infrared spectroscopy

When a substance is exposed to a mid-infrared energy source, its molecular structure is affected. Rather than causing a change in the distribution of electrons as occurs during a high energy x-ray exposure, mid-infrared exposure causes a change in the physical configuration of molecules (**Figure 2**). The molecular bonds bend and stretch as atoms are displaced from their normal-state equilibrium positions by mid-infrared energy. This results in vibrations along atomic bonds as they are affected by the radiation. The frequency of the vibrations, the amount and direction of atomic displacement, and the wavelength of energy that produces the reaction is unique to different substances according to their atomic constituents and

arrangement (Alpert and others, 1970). This relationship–that interactions with mid-infrared radiation are unique to certain materials–is the basis of field spectroscopy.

Figure 3, from Gaffey and others (1993), shows the four primary categories of bending and stretching as represented by linear, bent, and planar molecules such as carbon dioxide, water, and carbonate. **Figure 3Aa** depicts a linear molecule stretching as the peripheral atoms are displaced horizontally away from a central atom. In **Figure 3Ab**, bending occurs when the peripheral atoms are displaced vertically away from a central atom. Bending of peripheral atoms away from the center but in the z-direction (coming out of the page) is shown in **Figure 3Ac**. Finally, **Figure 3Ad** depicts stretching the atomic bonds horizontally so that one atom nears the central atom as the other pulls further away. Variations of these four types of bending and stretching are pictured in **Figures 3B** and **3C** for two other types of molecules. Modifications of bond activity are due to variations in the complexity of the atomic configuration, and the types of bonds and atoms present (Gaffey and others, 1993; Flett, 1970; Alpert and others, 1970).

The bond interactions described above are produced only by certain wavelengths of electromagnetic radiation within the mid-infrared. The vibrational modifications produced by bending and stretching of bonds require energy, which a material absorbs from the incident mid-infrared radiation. The quantity of energy absorbed is proportional to the displacement of atoms and subsequent vibrational modifications are inherent to specific chemical configurations (Gaffey and others, 1993; Flett, 1970; Alpert and others, 1970).

In order to measure the amount of infrared energy absorbed by a substance, high resolution instrumentation has been developed that collects and accurately records reflected energy at each wavelength. Identification of minerals using such instrumentation often concentrates on characteristic overtones of fundamental vibrational modes (Spectral International, 1994a; Gaffey and others, 1993; Jones and Childers, 1993). The concepts of "overtone" and "fundamental mode" are equal between spectroscopy and the physics of a vibrating string. Overtones are sometimes used in spectroscopy because significant features can be found when the atoms in a molecule do not move in phase along a straight-line path to define a fundamental mode. When excited by infrared energy, some molecules will experience Lissajous movement which follows more complicated paths of vibration with higher freqency than a fundamental mode. The more intricate motions produce overtones of the fundamental frequency of vibration. The overtones can be resolved mathematically into their fundamental components, and the wavelengths of their occurrence calculated (Gaffey and others, 1993). Whatever energy is not absorbed by the vibrations is reflected. The patterns of reflected energy are configured as graphical plots,

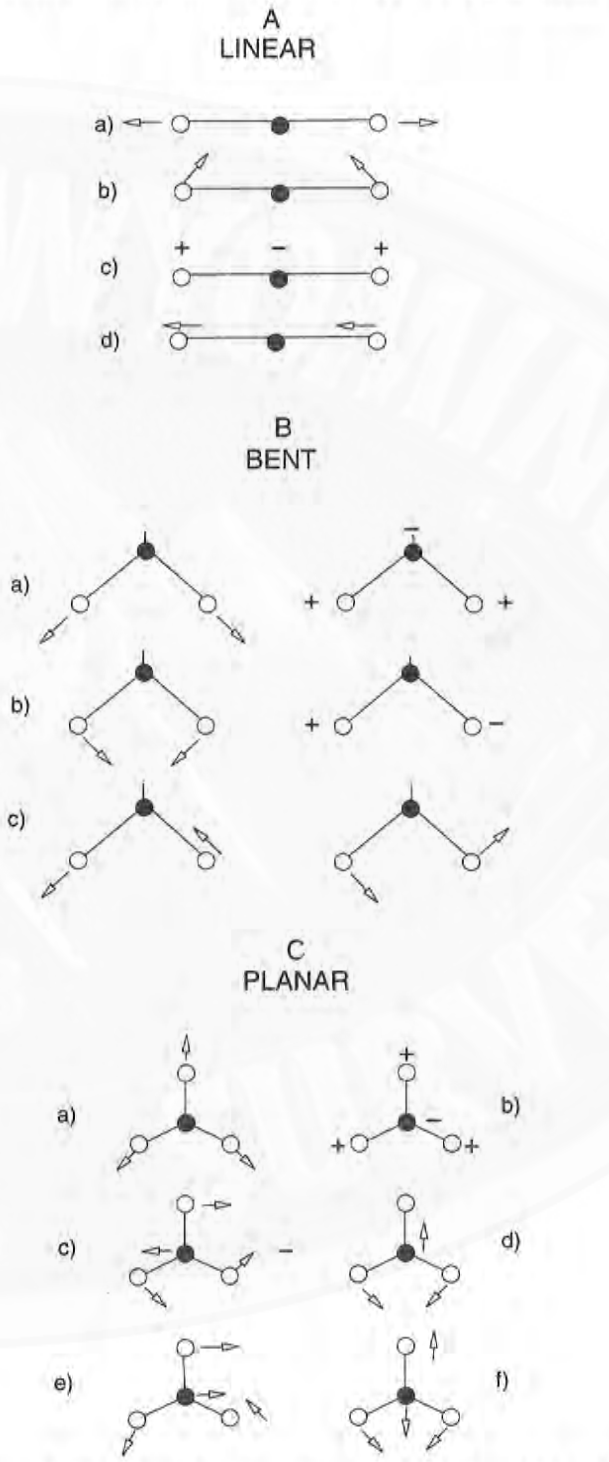


Figure 3. Three basic atomic configurations (A, B, and C) and the expected modifications made when exposed to infrared energy. Directions of displacement are shown by arrows on the page, +s out of the page, and -s into the page. (Adapted from Gaffey and others, 1993.)

called spectral curves, which are the interpretable results of a spectral measurement.

An example of a spectral curve is pictured in **Figure 4** and shows how much infrared energy was reflected from a sample at each wavelength between 1300 and 2500 nanometers. The 2-nanometer sampling interval used by the spectrometer producing this curve allows very narrow spectral features to be described for a sample. The dolomite sample represented by this curve shows a uniform reflectance between 1300 and 1830 nanometers indicating very little molecular response in this region. The pronounced dips beginning at 1900 and 2200 nanometers are the absorption features unique to the sample, indicating the basic carbonate molecular structure is responding in this spectral range. It is at these wavelengths that characteristic bending and stretching takes place in the carbonate molecule, absorbing the energy. The recorded molecular interaction thus defines dolomite spectrally, making this curve a "signature" for dolomite.

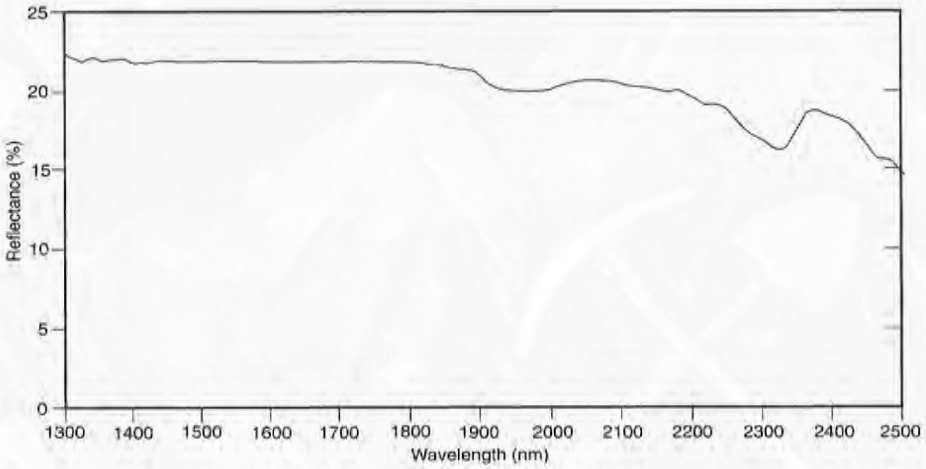


Figure 4. Spectral curve for dolomite showing percent reflectance in the mid-infrared (Sample C-05A, JPL Spectral Library), National Aeronautics and Space Administration/Jet Propulsion Laboratory (1992).

Dolomite is only one of the many mineral species that exhibit definitive spectral characteristics in the mid-infrared range. In addition to dolomite, other carbonates, amphiboles, clays, evaporites, hydroxides, micas, phosphates, sulfates, arsenates, feldspathoids, organic materials, silicates, and zeolites are also identifiable using infrared spectroscopy (Spectral International, 1994d; Hunt and Salisbury, 1970, 1971).

Numerous examples indicate the practicality of spectroscopy as a tool for spectral differentiation of materials based on those features that are related to or caused by molecular configurations (Gaffey and others, 1993; Hauff and others, 1991; Hunt and Salisbury, 1970, 1971). Laboratory spectroscopy has been used for some time as a method for discriminating various minerals, but its application, like that of XRD, is limited due to the difficulties inherent in acquiring representative samples, transporting them to a lab, and awaiting the results. In this respect, laboratory spectroscopy is comparable to XRD. Infrared spectroscopy becomes a superior identification method when its application is accomplished using hand-held instrumentation in the field, a feature thus far impractical for XRD analyses.

Instrumentation

Companies such as Spectral International, Inc. (Lafayette, Colorado) and Analytical Spectral Devices, Inc. (Boulder, Colorado), among others, have visible range and infrared spectrometers on the market that cater specifically to scientists needing to measure spectral responses in the field. As an example, Spectral International currently sells a shoebox-sized spectrometer that reads short wave infrared energy (SWIR). Its general workings and design are pictured in **Figure 5**.

This particular spectrometer, as described in the PIMA II™ Operations Manual (Spectral International, 1994e) shows its truly portable nature made possible through recent technological developments. It has a weather-proofed outer casing and an internal halogen light source so that results will not be affected by variations in solar illumination or atmospheric water interference. The instrument weighs only 5.5 pounds and can be powered by batteries or plugged into an automobile cigarette lighter. Up to 600 measured spectra can be stored in the tiny palm-top computer integrated into the unit. The computer allows results to be viewed immediately on a high resolution liquid crystal display (LCD) screen or downloaded later to an external personal computer (PC) for analysis, manipulation, and print-out. Measurements take 20 to 30 seconds for each sample and realize a 6- to 10-nanometer spectral resolution, depending on whether a 2- or 4-nanometer sampling interval is used.

The point-and-shoot character of PIMA II™'s contact data collection is facilitated by a sophisticated

internal design (**Figure 5**). Its basic function begins when the halogen light source (A) emits infrared energy in wavelengths between 1300 and 2500 nanometers. The infrared beam (B) is directed through a diffraction grating (C) towards an external lens (D) to strike a sample surface. The energy effects a vibration within the sample as described earlier, and the first or second overtones of the fundamental mode of the motion are read by receptors (E) within the instrument. The response is standardized using an internal calibration system and then digitized to appear graphically in a reflectance plot (**Figure 4**) on the palm-top computer screen (F). Spectra are available for visual analysis in the field as soon as they are collected, and can be readily transferred to another computer where results may be mathematically detailed using system software installed on an office PC.

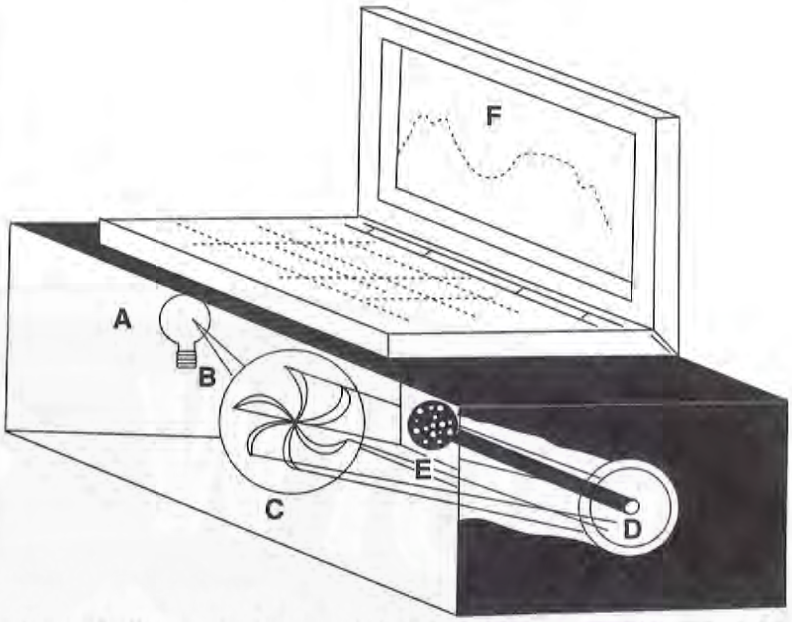
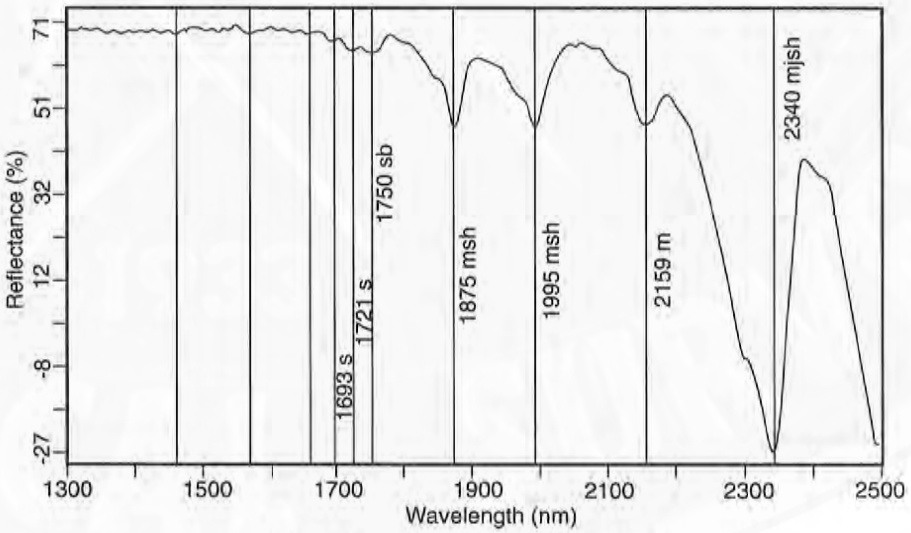


Figure 5. Schematic drawing of the general workings within a portable spectrometer. See text for details and component parts. (Modelled by C. Jenkerson after PIMA II™ manufactured by Integrated Spectronics, Pty.)

Analytical methods

Reading, understanding, and interpreting data from a field spectrometer appears confusing and tedious at first (**Figure 6**), but is easily accomplished with practice and study. Although formats have not yet been formally standardized, there are three basic recognizable properties on every spectral curve that are used to distinguish materials in the SWIR. The curves are typically described according to the position, intensity, and general shape of their absorption features (Spectral International, 1994c). Position is the location of a feature along the x-axis, defining at which wavelength the feature occurs. Intensity is displayed along the y-axis, showing what proportion of the source energy was reflected from the sample. The third property is given by a somewhat subjective determination of the general shape of a feature (i.e., broad, sharp, major, symmetrical).

A description of a spectral curve is exemplified in **Figure 6** by a curve generated for a pure calcite



POSITION	INTENSITY	SHAPE
1693	03%	small
1721	05%	small
1750	05%	small, broad
1875	20%	medium, sharp
1995	20%	medium, sharp
2159	11%	medium
2340	65%	major, sharp

Figure 6. Interpretation of absorption features on a spectral curve for pure calcite. The position, intensity, and shape of a feature are diagnostic in mineral identification. (Adapted from Spectral International, 1994d, section 3.)

sample. Note that the position of significant features is marked in nanometers of wavelength, the intensity is related to the percent reflectance, and the initials marked at each feature designate the shape of the features as small (s), medium (m), or major-deep (mjd).

Even with the general subjectivity involved with labeling feature shapes, spectral curves such as this have been used effectively, particularly in the identification of minerals (Spectral International, 1994c, 1994d; Burns, 1993; Gaffey and others, 1993; Hauff and others, 1991; Hunt, 1977; Hunt and Salisbury, 1970, 1971). To an experienced spectral worker, subtle differences in the features can indicate differences between and within materials. Even slight variances from the ideal curve for a particular mineral, for example, can signify important genetic properties like temperature, acidity, depth of formation, or crystal development (Hauff, 1994). Spectral analysis can thus be effectively used not only for identification, but to track crystallinity and elemental substitutions, to map alteration intensities and alteration zone variation, and even to discriminate between altered and weathered materials (Hauff, 1994; Spectral International, 1994b). For an exploration or environmental geologist, field spectroscopy can thus provide an interpretative understanding of a mineral environment in a fraction of the time and at a lower overall cost than traditional sampling and laboratory methods.

Application examples

Clays

Clay minerals commonly indicate the temperatures, pressures, or the chemical environment of a geologic site, and are very amenable to infrared analysis due to the sensitivity of infrared to the waters and hydroxides present in most clays (Gaffey and others, 1993). What is particularly useful about infrared spectroscopy in clay research is that it enables the distinction between clays that appear ambiguous using other identification methods (Hauff, 1994).

Figure 7 depicts the spectral signature of several common alteration minerals. Of note are curves C and D, the infrared responses of a dickite and a kaolinite. They are similar minerals chemically, which is sometimes annoyingly obvious in the field, and the spectral curve reflects this. But the spectral curve also shows absorption features that easily distinguish the two closely related clays.

The largest distinction is apparent at 1400 nanometers. Here, a deep "trough" displays "shoulders" topping its left and right limbs, and a "doublet" at the minima (creating a "W" shape). The shoulders are the kinks that step into and out of the trough, and are more recognizable in the kaolinite (curve D). The left shoulder on curve D is sharp and narrow while the right shoulder is broad and gentle. Shoulders on the dickite (curve C) are almost indefinable by comparison. The second difference between the dickite and kaolinite in this wavelength position is that the doublet is easier to interpret. The dickite doublet clearly spans more percentage points of reflectance than the kaolinite doublet, giving it a greater intensity and a more pronounced shape.

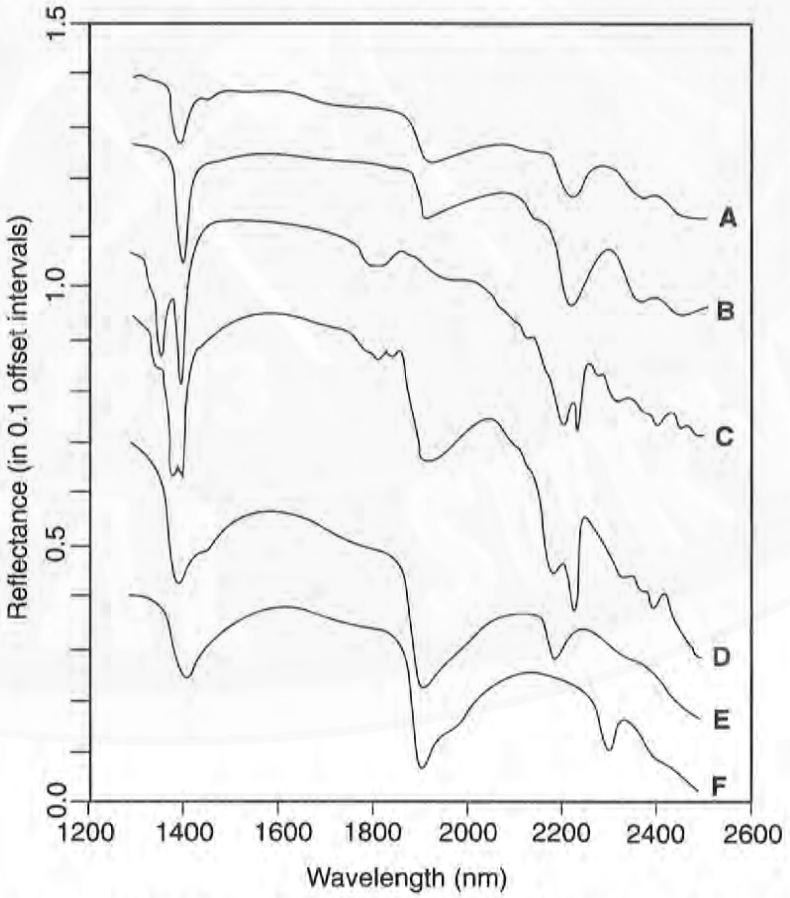


Figure 7. Comparison between spectral curves of six common alteration minerals: (A) illite, Nevada; (B) illite, Arizona; (C) dickite, Australia; (D) kaolinite, Nevada; (E) beidellite, Idaho; (F) nontronite, California. (Adapted from Hauff, 1994.)

A second absorption feature is displayed in these two curves and although it is apparent on this graph, it is not always a characteristic mark (Spectral International, 1994d). The difference here is in the intensity, shape, and position of the trough sequence between 1800 and 2100 nanometers. The first dip has a more abrupt introduction on the dickite curve (C), and has a smoother bottom. Starting around 1900 nanometers, the more obvious difference is that the second dip on the kaolinite curve is much deeper and better defined than the lazy sag on the dickite curve.

Finally, a close look at the doublet in each curve above 2200 nanometers shows another distinguishing feature. The doublet in the dickite (at 1400 nanometers) was the more intense, but at 2200 nanometers, the kaolinite feature is slightly stronger and sharper, particularly along its right limb.

The distinctive absorption features at 1400, 1900, and 2200 nanometers that designate dickite from kaolinite are also present in the other plots on **Figure 7**, and a brief perusal of their characteristics will show how those representative alteration minerals can be distinguished from one another. In fact, those three wavelengths (1400, 1900, 2200 nanometers) are common positions for the strongest features in clay spectra. Not surprisingly, these positions have been proven chemically to represent the major absorption bands for water and hydroxides (Spectral International, 1994c).

Because certain hydroxide compounds are best defined at one wavelength or another, clays containing them can also be characterized according to their responses at that wavelength. The spectral differences between illite and bentonite are good examples of this (**Figure 8**). One way to compare the two is by concentrating on the differences between them based on illite being an aluminum hydroxide clay and bentonite being a silicon hydroxide clay. Aluminum hydroxide (AlOH) absorbs infrared strongly around 2200 nanometers (Spectral International, 1994c). As seen on **Figure 8**, the illite curves show better defined features at this wavelength, collaborating the characteristics shown spectrally by aluminum hydroxide. Likewise, the strong response of silicon hydroxide (SiOH) around 1900 nanometers exaggerates bentonite's curve beyond what is shown in the illite (Spectral International, 1994c, 1994d).

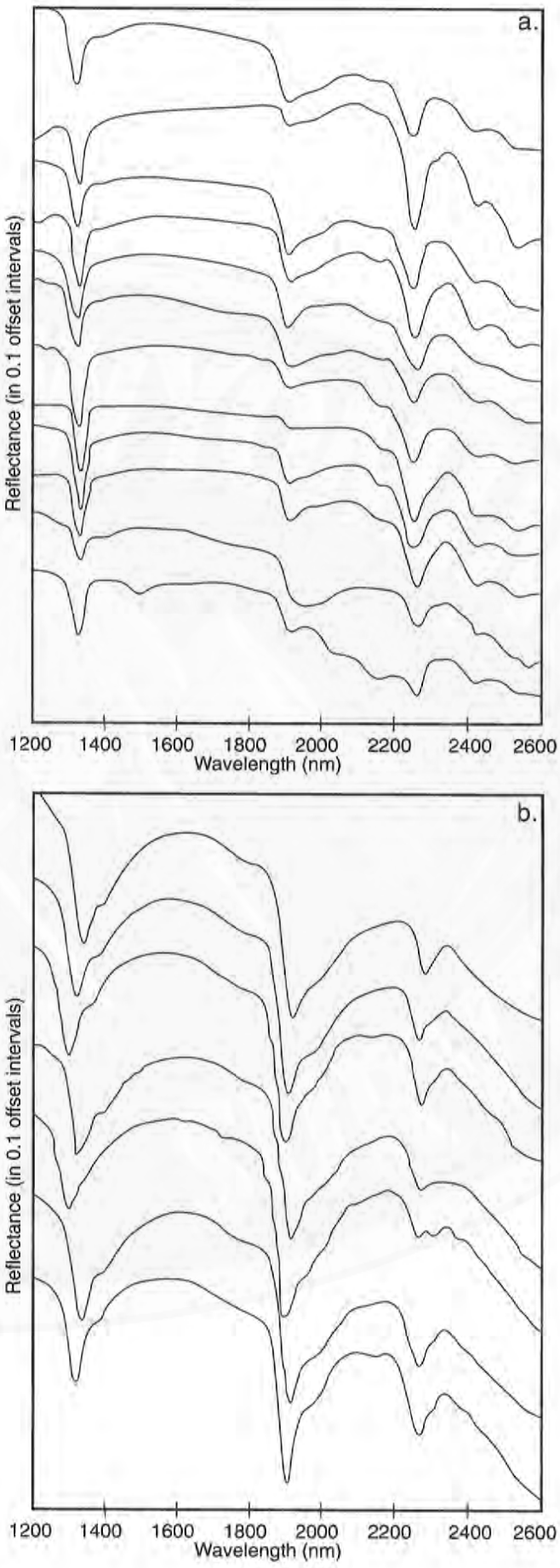


Figure 8. Reference spectra for a) illite and b) bentonite, showing typical variation within and comparison between clay types. (Adapted from Spectral International, 1994d, section 4.)

Kimberlites

Spectral distinctions such as those described above between dickite and kaolinite, or illite and bentonite, represent the viability of field spectroscopy for clay mineral studies. This has been found useful for mapping hydrothermal alteration zones and kimberlite intrusions, where characterization of surficial materials can quickly outline exploration sites (Spectral International, 1994b).

Minerals associated with kimberlitic environments are under continuous study. Currently, those minerals are expected to be amenable to infrared response measurement (Spectral International, personal communication, 1993), and the potential presented by field spectroscopy may help speed site assessments and reduce laboratory analyses. Like hydroxides, the mineral assemblages associated with and weathered from the kimberlitic diatremes have distinctive absorption band patterns. Differences between kimberlitic clays and surrounding materials can often be easily distinguished by a field spectrometer. The chromium (Cr), calcium (Ca), and magnesium (Mg) content in indicator minerals, whose proportions may be related to diamond occurrence in kimberlites (Gurney, 1984), are good targets for spectral study.

Figure 9 depicts a recent measurement plotting a diopside crystal against its chromium-rich counterpart. The result is interesting, as the two curves seem to be mirror reflections of each other, with a greater general intensity found in the chromium sample. The reasons behind this occurrence are assumed to be chemical, and further study will determine its potential as a distinctive property.

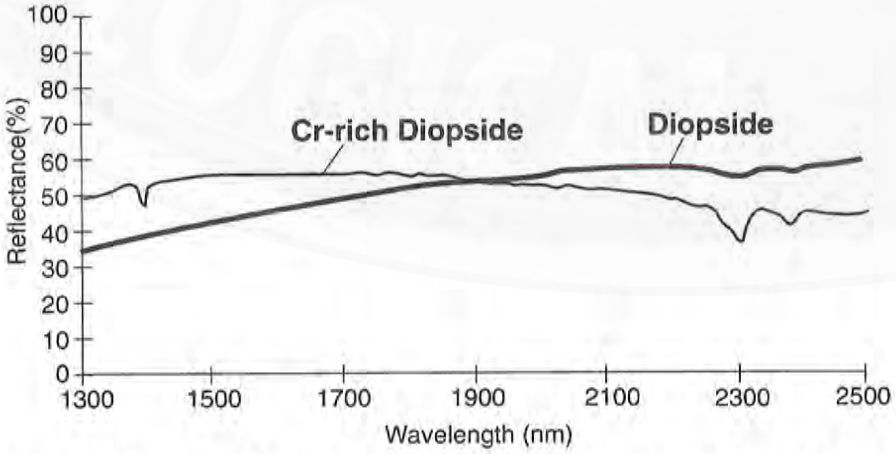


Figure 9. Comparison between two diopside samples. Chromium is thought to be a good indicator of kimberlitic terrain characteristcs (Gurney, 1984), and provides a good spectral target in certain minerals.

Zeolites

Another mineral species of interest is the zeolites. Again, spectral analysis of zeolites is expected to provide researchers with a valuable tool for assuring otherwise difficult identifications (Spectral International, personal communication, 1993). The application of infrared energy to zeolite mapping is just becoming a priority in spectral research, but some data is available.

For example, **Figures 10** and **11** present spectral signatures of several types of zeolites. **Figure 10** shows formal measurements from a reference library (Clark and others, 1993). Each of the curves is similar, as might be expected, but position offsets, magnitudes of intensity, and general shapes of the absorption features make most of them distinguishable from one another. These differences may seem subtle, but in fact can be enough to make important determinations of identity when mapping in the field.

Two referenced plots are shown with raw data collected for unknown zeolite samples (**Figure 11**). (NOTE: The "noise" visible in the sample plots is due to an unfortunate malfunction in the spectrometer used for this measurement. The reference curves for clinoptilolite and chabazite are better representatives of a spectrometer's capabilities.) Comparison between absorption features suggests that both large and small samples of "white tuff" are more likely to be clinoptilolite than chabazite. They each lack the small but distinct shoulder on the right limb of the trough at about 1450 nanometers, and the more gradual rise between 2100 and 2300 nanometers found in the chabazite reference curve is not present in the white tuff samples. In addition, the sharper incline into and out of the trough at 1900 nanometers is better matched to that in the clinoptilolite.

Studies by several workers are currently building spectral libraries for zeolites that will soon provide an essential reference set for identification (Spectral International, 1994d). As with kimberlite studies, interest and collection of data is expanding so that infrared measurements can be applied to zeolites, and to other less common minerals and surficial materials.

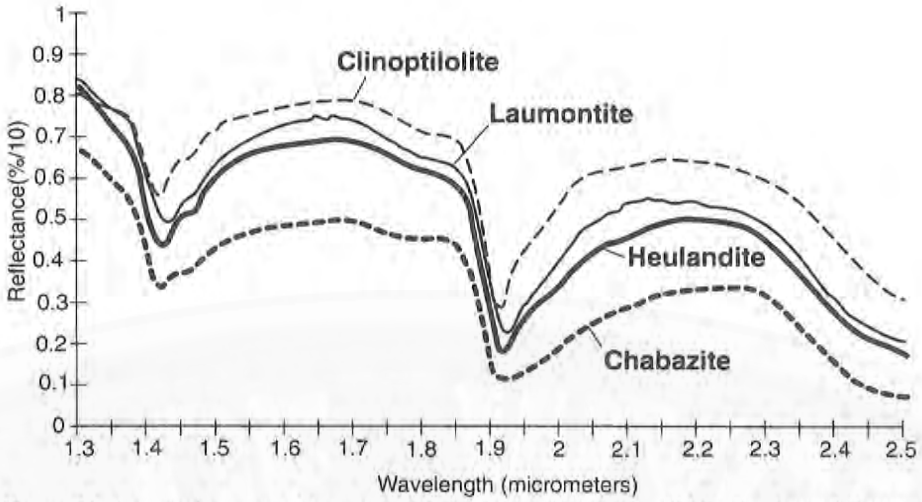


Figure 10. Spectral curves of four minerals in the zeolite species (U.S. Geological Survey Digital Spectral Library, Clark and others, 1993).

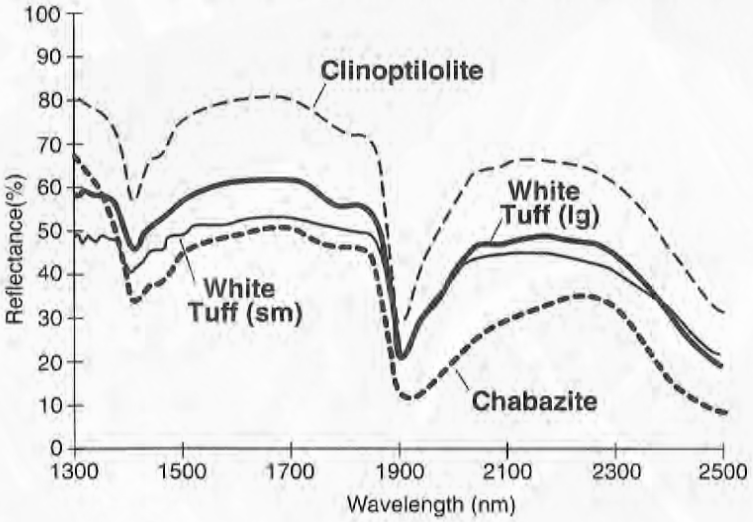


Figure 11. Spectral measurements of two unknown zeolites from the Washakie Basin, Wyoming (courtesy of R. Harris, Wyoming State Geological Survey) plotted with two referenced zeolites from the U.S. Geological Survey Digital Spectral Library (Clark and others, 1993).

Summary

Spectroscopy is rapidly gaining a place in field-oriented sciences since it can be used to efficiently identify a great number of compounds, and as the instrumentation becomes more advanced and easier to use.

Infrared field spectroscopy is already considered essential to verification of remote sensing interpretations, and a worthwhile mapping tool that can quickly provide extremely valuable field data for applications in industrial mineralogy involving clays in particular. When compared to traditional field surveys, spectroscopy is relatively inexpensive and minimizes uncertainties in the identification of surface materials while reducing data processing time. This, and the reliability of spectral identification, make spectroscopy a valuable asset for mapping, exploration, or monitoring surficial conditions.

Acknowledgments

The authors acknowledge with gratitude Mr. Ray Harris of the Wyoming State Geological Survey, for inviting this paper and providing zeolite samples. Thanks also to Dr. Ronald Marrs for his advice and commentary. Funding from the Wyoming Space Grant Consortium (Planetary and Space Science Center) and the Department of Geology and Geophysics at the University of Wyoming is greatly appreciated. Finally, some sample spectra were obtained from Phoebe Hauff, Spectral International, Inc. (Colorado), and from the U.S. Geological Survey and National Aeronautics and Space Administration/Jet Propulsion Laboratory Spectral Libraries.

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U.S. Geological Survey Digital Spectral Library
http://speclab.cr.usgs.gov/spectral-lib.html



Outcrop of zeolite-bearing tuff bed, Washakie Formation, Washakie Basin, southwestern Wyoming. Top of rock hammer (circled) indicates approximate top of white tuff bed. (Photograph by R.E. Harris.)

Diamonds in Virginia

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Abstract

There are four sites in Virginia and one just across the state line in West Virginia where diamonds have been found; two kimberlite sites in Virginia (a spinel peridotite and a mica peridotite) may have some potential for diamonds. Diamond sites include the Vaucluse gold mine in Orange County; the Whitehall gold mine in Spotsylvania County; the Gillespie farm near Pounding Mill in Tazewell County; the "Dewey Diamond" at Manchester (now south Richmond); and the "Punch Jones Diamond" in Peterstown, West Virginia. All five diamonds were found near or adjacent to large river drainages. The two kimberlite sites are the Mount Horeb kimberlite, located southwest of Lexington in Rockbridge County of west-central Virginia, and a mica peridotite, which occurs west of Front Royal, Warren County, northern Virginia.

There are three main sources of diamonds: kimberlites, lamprophyres (mafic to ultramafic igneous rocks), and alluvial or placer deposits derived from weathering of these igneous rocks. The most common diamond source in Virginia is in alluvial deposits where diamonds have been removed from their bedrock source. The bedrock may be kimberlite, which is usually dark green in color and composed mainly of the silicate minerals olivine, serpentine, mica, and some pyroxene or the bedrock may be lamprophyres, which are dark dike rocks containing biotite, hornblende, or pyroxene with the dark minerals occurring as phenocrysts and in the groundmass. Kimberlites and lamprophyres occur in a belt from Alabama through Virginia to eastern Canada.

Many other basic and ultrabasic igneous rocks are found in the western Valley and Ridge and Blue Ridge provinces of Virginia. Many are in Augusta, Bath, Highland, and Rockingham Counties in the central western part of the Commonwealth of Virginia.

Introduction

Five diamonds have been found in or near Virginia; two additional sites may offer some potential for diamond exploration (**Figure 1**). The five diamonds were found in the City of Richmond; in Orange, Spotsylvania, and Tazewell Counties, Virginia; and in Monroe County, West Virginia. Holden (1944a) noted that Stephenson (1878) reported that diamonds

> *. . . several of great value have been accidentally found in Virginia in former years. One, of more than forty carats, was found in a farmer's desk, after his death, which had lain there for more than a quarter of a century without his knowing anything of its value.*

The Mount Horeb kimberlite in Rockbridge County and a mica peridotite dike in Warren County may offer some economic potential. There are many other ultrabasic igneous rocks in the western part of Virginia, but they are not of kimberlitic origin.

Sources

Of the three sources of diamonds (kimberlites, lamprophyres, and placer deposits), kimberlites are the primary source. They are derived from the melting of rocks that compose the earth's mantle. Kimberlites are usually dark green in color and composed mainly of the silicate minerals olivine, serpentine, mica, and

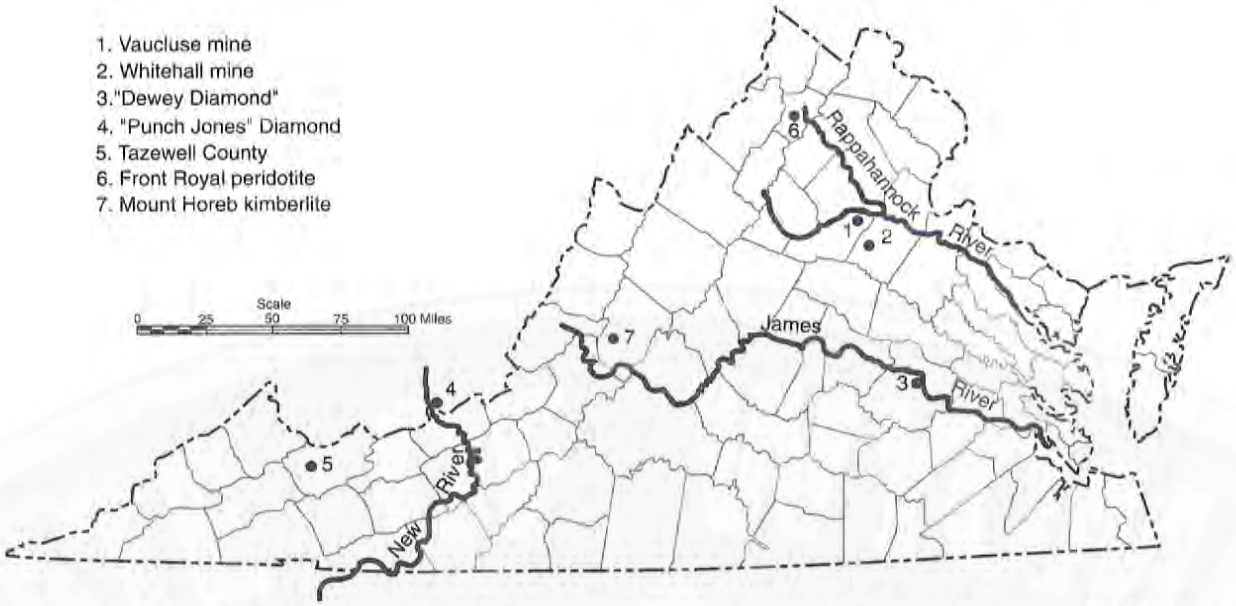


Figure 1. Location of diamonds and potential exploration sites (kimberlite and peridotite occurrences) in Virginia.

some pyroxene; they also contain calcite and iron-titanium oxide minerals and are found in geologic settings where the earth has been tectonically inactive for millions of years (Pasteris, 1983). Kimberlite is produced when peridotite (dark intrusive rock) melts at different levels in the earth's upper mantle. Initial melting may occur at depths of 150 to 200 km where temperatures are over 800 C and pressures are greater than one million pounds per square inch. The right combination of great pressure and (relatively) low temperature is called the diamond-stability field (Bruce, 1995). The peridotite melt is practically a liquid and allows carbon (C) atoms to freely move around and form into diamonds. Dating of zircons, which are commonly found as inclusions in diamonds, provides a Precambrian age (older than 550 million years) and indicates that the diamonds have been flowing around in the mantle for a very long time (Sweet, 1994).

Major diamond deposits occur in "pipes" or diatremes that branch toward the surface (Pasteris, 1983). A diatreme may consist of several intrusions of kimberlite and may have erupted like a volcano. As the kimberlite melt rises, it captures fragments of surrounding rocks, which are carried along with the intrusive; crystallization also takes place nearer the surface and changes the chemistry of the melt. The end product may be entirely different from the composition of the original melt in the mantle. Although kimberlites are the main source rock for diamonds, only about five percent of known kimberlites contain diamonds. The richest kimberlites thus far discovered contain about 20 to 30 carats of diamonds per 100 tons of rock (Sweet, 1994).

In exploration, indicator minerals that are usually eroded from the kimberlite include reddish pyrope garnet, dull-hued chromite, iron-heavy ilmenite, and bright green diopside. During exploration in the Northwest Territories of Canada in the late 1970s, garnets, chromite, and other indicator minerals with subtly different chemical signatures were determined to be present in diamond-bearing kimberlites but absent in barren kimberlites (Krajick, 1994). These minerals included G10 garnets, which are lower in calcium (under 4%) and higher in chromium than other garnets, and diamond-inclusion chromite which is higher in chrome (62.5% or more) than other garnets. When iron in ilmenite was highly oxidized, a kimberlite contained virtually no diamonds. It was determined that iron took on more or less oxygen in response to heat and available oxygen in the magma. When iron became highly oxidized, so did the diamonds, essentially vaporizing into carbon dioxide. As a result, G10 garnets and chromite determine how many diamonds entered the kimberlite and the ilmenite determines how many survived (Bruce, 1995).

A second diamond source includes mafic and ultramafic rocks, such as lamprophyres, present in a belt of rocks from Alabama to eastern Canada. A third source is the alluvial or placer type diamonds which were deposited by streams or rivers that drain regions where kimberlites or lamprophyres occur. Gravels weathered from these types of rocks have yielded diamonds in Virginia and several other eastern states (Wilson, 1948). The specific source of these diamonds is still unknown, as it is usually difficult to trace a given deposit to its bedrock source. Gold (1984) stated that of the 300 tons (1.5 billion carats) of diamonds recovered to date, most have come from alluvial workings.

Virginia diamonds

There have been five diamonds found in or near Virginia and there are at least two kimberlites that warrant further exploration. The five diamonds are: the "Dewey Diamond," found in what is present-day Richmond; a diamond from the Vaucluse gold mine in Orange County; a diamond from the Whitehall gold mine in Spotsylvania County; a diamond from a garden near Pounding Mill in Tazewell County; and the "Punch Jones" diamond found near Rich Creek, in Peterstown, West Virginia. The two sites that may offer some economic potential are the Mount Horeb kimberlite in Rockbridge County and a mica peridotite dike in Warren County. There are many other ultrabasic igneous rocks in the western part of Virginia, but they are not of kimberlitic origin.



Figure 2. Northeast corner of Ninth and Perry Streets, in the city of Richmond, where the "Dewey Diamond" was found in 1854.

Dewey Diamond - city of Richmond

The "Dewey Diamond" (Manchester Diamond), was found at the southwestern corner of Ninth and Perry Streets in Manchester (now a part of the city of Richmond), in the floodplain of the James River, by Benjamin Moore, a laborer (**Figure 2**). Mr. Moore was leveling a hill at this location in 1854 and found the 23.75-carat diamond in the soil (Green, 1982). M.C. Gilbert (1982, written communication) states that this diamond could have originated from the Mount Horeb kimberlite in Rockbridge County, as it is in the James River drainage basin.

The diamond was taken to a jeweler in Richmond, across the James River, and appraised at a value of $4000. However, Mr. Moore sold it to Captain Samuel W. Dewey, a geologist and mineralogist from Philadelphia for $1500. The first account of the diamond was published in the April 18, 1855 issue of the *New York Evening Post*, where it was called the "Dewey Diamond" after Captain Dewey and was described as the "largest diamond found in North America." Dietrich (1990) stated that the diamond was also referred to as the "Manchester" diamond. As noted by Kunz (1890), the diamond was described by John H. Tyler as an octahedron with slightly rounded faces and having a large flaw on one side. Watson (1907) described the color of the diamond as "a faint greenish-white with perfect transparency, but the refraction was somewhat impaired by a flaw or speck in the interior." Green (1982) noted a newspaper account from 1935 describing the stone as "a rather pretty pebble about the size of a hickory nut." In 1855, Dewey, attempting to eliminate the imperfection, had the diamond cut by the firm of H.D. Morse in New York City at a cost of $1500, which reduced its size from the original weight of 23.75 carats to 11.15 carats. Sinkankas (1959) noted that "when cut, the gem proved off-color and imperfect and was worth considerably less than the cost of cutting alone." Watson (1907), stated the weight of the cut diamond to be 11 11/16 (11.69) carats. Later, probably in the late 1850s, the diamond was sold to John Morrisey, a prize fighter in the days of bare fists, for $6000. Morrisey, who became the American heavyweight champion in 1858 when he defeated John Heenan, had the diamond set in a ring. He retired from fighting and operated luxurious gambling houses in New York; he also served two terms in Congress (Green, 1982). Morrisey, as the last owner of the diamond on public record, died in the late 19th century. It was reported that exact glass copies of the Manchester Diamond, as it was found and as cut, are in the U.S. Mint Museum in Philadelphia, Pennsylvania and at the Peabody Museum of Natural History in New Haven, Connecticut (Watson, 1907). Telephone calls to both museums in 1995 were fruitless as no glass copies were on display and no records were available. Green (1982) stated that a glass copy of the stone is also in the Smithsonian Institution in Washington, D.C. (**Figure 3**). According to Holden (1944a), there is also a model of the diamond in the Field Museum of Natural History, Chicago, Illinois.

Vaucluse mine – Orange County

In eastern Orange County, north of the end of State Road 667, is the site of the abandoned Vaucluse (Orange-Grove) gold mine where a "diamond of the first water" was reported to have been found in gold washings in 1836 (**Figure 4**). The same reference noted that since there was no "lapidary at hand," the diamond's real value was never ascertained (Anonymous, 1847a). In 1847, the diamond reportedly remained in the possession of the proprietors of the



Figure 3. Glass copy of the "Dewey Diamond" at the Smithsonian Institution, Washington, D.C. (photograph courtesy of the Smithsonian Institution).



Figure 4. Flooded drainage where a diamond was found "in washings" in 1836 at the Vaucluse gold mine in Orange County.

mine (who were Thomas H. Boswell and William H. White in March, 1835) at the time (Anonymous, 1847b).

Whitehall mine – Spotsylvania County

Stephenson (1878) noted that Mr. G.W. Featherstonhaugh, an English geologist, found a diamond at the Whitehall mine in west-central Spotsylvania County, north of State Road 608. Mr. Featherstonhaugh was also the editor of the Geological Magazine of Philadelphia (The Virginia Herald, 1833). Holden (1944) stated that Stephenson (who found a diamond in Burke County, North Carolina after 1843) was an assayer at the U.S. Mint at Dahlonega, Georgia and had considerable experience in placer gold mining.

Tazewell County

In 1913, a diamond was found by Frank Brewster, a laborer, in a cornfield 200 feet above a local stream bed, in Tazewell County, on the farm of J.S. "Straws" Gillespie. The site is located about 3 miles east-northeast of the intersection of Pounding Mill Branch Road (State Road 637) with U.S. Highway 460 in Pounding Mill (**Figure 5**). Mr. Gillespie's daughter Ollie was in the cornfield at the time and said that the stone was rough and unimpressive (O.M. Gillespie, 1995, personal communication). Mr. Brewster carried the stone around in his pocket for about a year before he sold it to H.W. Pobst, a jeweler in Tazewell, Virginia, who had it cut by J.R. Wood and Sons of New York. H.W. Pobst's son Tom said that the stone was a beautiful blue color and also noted that Wood had declared it to be a genuine diamond. Mr. Gillespie later acquired the diamond back before he passed away around 1920. According to Holden (1944), who described the diamond as a blue-white stone, no more stones had been found at this site in the following 30 years of searching. Sinkankas (1959) stated that the diamond was cut into a gemstone that weighed 0.83 carat. The



Figure 5. Site where a diamond was found on the J.S. Gillespie farm in 1913, near Pounding Mill, Tazewell County.

diamond has now been set in a ring and remains in the Gillespie family (O.M. Gillespie, 1995, personal communication).

"Punch" Jones Diamond – Monroe County

In April, 1928, a diamond that was even larger than the Dewey Diamond was discovered (**Figure 6**). This diamond known as the "Punch" Jones Diamond was found along Rich Creek, in Peterstown, West Virginia, less than one thousand feet from the Virginia state line. It is believed that the diamond was actually from a source in Virginia, North Carolina, or Tennessee and washed down the New River into Rich Creek and then into West Virginia. Holden (1944a, b) noted that due to the abundance of "carry impact markings" and the size of the stone, it had probably traveled a long and violent journey. The diamond measured 5/8 inch across with 12 diamond-shaped faces, weighed 34.48



Figure 6. Historical marker for the "Punch" Jones Diamond along Rich Creek, Peterstown, West Virginia.



Figure 7. Location of the site of the horseshoe pit, beside the house, where the "Punch" Jones Diamond was found in 1928, in Peterstown, West Virginia.

carats (Holden, 1944b, noted the weight as 34.46 carats), and was the largest found in North America at the time of its discovery.

The bluish-white diamond was found by William P. "Punch" Jones and Grover C. Jones, Sr., while pitching horseshoes in their yard in April, 1928. "Punch" noticed a bright object when a horseshoe rang in the pit; he picked up a rounded, "glassy mass of striking brilliance, about 3/4 inch in diameter" (**Figure 7**). In 1942, "Punch" Jones, then 26 years old, asked his father about the stone that they had found fourteen years earlier. It had been stored in a wooden cigar box, stashed in a toolshed (**Figure 8**). After "Punch" convinced himself the diamond was genuine, he contacted Roy J. Holden, a geology professor at Virginia Polytechnic Institute in Blacksburg on May 5, 1943. After Holden authenticated the diamond, it was sent to the Smithsonian Institution by Jones for safekeeping and remained there until early 1964. "Punch" Jones was killed in World War II and possession of the diamond passed to his father. In February, 1964, the Joneses brought the diamond home and put it in a bank deposit box in the First Valley National Bank in Rich Creek, West Virginia.



Figure 8. Toolshed where the "Punch" Jones Diamond was "stored" for fourteen years.

When Grover Jones died in 1976, his widow Grace and grandson Robert became owners of the diamond. In 1984, Robert sold the diamond through Sotheby's auction house in New York to an agent representing a lawyer in the Orient for $74,250 (Macy, 1990). Grace Jones passed away in 1992.

Kimberlites in Virginia

Two known kimberlites in Virginia, a mica peridotite dike in Warren County and the Mount Horeb kimberlite in Rockbridge County, are in the eastern edge of the Appalachian Basin as noted by Southworth and others (1993). Parrish and Lavin (1982) note there is geologic evidence in western Pennsylvania for structure parallel (linear features parallel to the regional structure) Cambrian-Ordovician growth faults; this has been verified by aeromagnetic data. The Virginia kimberlites may be parallel to the regional Paleozoic faults although there is no clear evidence of basement faults. Northwest-trending dikes of Jurassic age are present in the Shenandoah Valley and crystal zones of weakness may be created under extensional stress conditions and reactivation of the strike-parallel faults at the intersection of the dikes, creating cross-strike faults.

The westernmost limit of these dikes appears to be a Late Jurassic mica pyroxenite dike in southern Pendleton County, West Virginia from which biotite produced an argon-argon (Ar-Ar) age of about 147 Ma (million years) (Southworth and others, 1993).

The undated mica peridotite dike in Warren County is similar to the Pendleton County dike. Garner (1956) noted that peridotite porphry dikes are located on U.S. Highway 33, about 13.5 miles west of the Virginia-West Virginia state line, and consist of 30% phenocrysts of augite and olivine in a groundmass of augite, magnetite, and biotite plates.

The Mount Horeb kimberlite, which is also undated, may exist with the Warren County kimberlite in a belt of kimberlite intrusives in the eastern edge of the Appalachian Basin, according to Parrish and Lavin (1982).

Mount Horeb - Rockbridge County

A kimberlite occurs in the northeastern portion of the Sugarloaf Mountain 7.5-minute Quadrangle, near Mount Horeb Church in western Rockbridge County. At the surface, the kimberlite consists of three separate intrusions into an Ordovician age carbonate sequence (Sears and Gilbert, 1976) (**Figure 9**). A geologic map of the area, published by Spencer (1968), shows the kimberlite intruding the Beekmantown Dolomite. Sears and Gilbert (1973) noted that unaltered primary crystals in the kimberlite include chrome-rich spinel, green clinopyroxene, pyrope garnet, mg-ilmenite, rutile, zircon, and amphibole. Altered material consists of montmorillonite, vermiculite, and chlorite. Xenolithic fragments of Middle Ordovician sandstone and fossiliferous limestone are also present. Steidtmann (1948) reported that the peridotite intrusion was in the Conococheague Limestone and that about one-half of the rock consists of phenocrysts (olivine ?) completely replaced by serpentine. He also noted that the groundmass contains doubly terminated apatite crystals, which occur as "swarms" around, but not in, the phenocrysts.



Figure 9. Exposure of weathered kimberlite off the east side of State Road 807, near Mount Horeb Church, Rockbridge County.

According to Sears and Gilbert (1973), the chemistry is sufficient to verify the body as a kimberlite and establish its mantle origin. However, Gilbert (1982, written communication) noted that since spinel is so much more abundant than garnet, a shallower depth of origin such as 50 to 100 km instead of the normal 150 to 200 km for kimberlite seems likely. The Mount Horeb body may be more aptly termed a spinel peridotite instead of a garnet peridotite.

A magnetic survey has been conducted up the drainage on the west side of the southernmost intrusion; an anomaly was noted near the road and well into the body to the east. Panning in the drainage has produced ilmenite, pyrope garnet, green diopside, and zircon (A.W. McThenia, 1995, personal communication). Some exploration has been done on sites of two of the intrusions, which are presently owned by two separate landowners. Cominco examined the kimberlite in the early 1980s and Kennecott Corporation was on the property in the late 1980s. No diamonds have reportedly been found at either of these sites.

In the fall of 1995, the Virginia Division of Mineral Resources drilled several auger holes off the northeastern side of State Road 807 (Pullen Road) and on the southwest slope of the kimberlite. The most successful hole was spudded at 1510 feet elevation in weathered kimberlite and was drilled to a depth of 34 feet; total depth remained in weathered kimberlite.

Mica peridotite - Warren County

A mica peridotite dike (600 feet x 150 feet) occurs in Warren County, 4.5 miles west of Front Royal, off the northwestern side of State Road 626, approximately 0.35 mile by road northeast of its intersection with State Road 617 (**Figure 10**).

From thin section analyses reported by Young and Bailey (1955), the minerals present were chlorite, 43%; phlogopite, 31%; hydrobiotite (?); pseudomorphs after olivine and pyroxene, 13%; secondary pyrite, 5%; perovskite, partly altered to leucoxene, 4%; apatite, 2%; dolomite, 1%; and ilmenite, magnetite, and epidote combined, less than 1%.

The dike, which has been chloritized, was examined by Rader and Biggs (1976). They made two magnetic traverses across the dike using a Geo-Metrics® model G-806 portable magnetometer. The traverses show a magnetic zone in the body of the dike about six feet thick. Weathered samples were obtained from three pits dug in the easternmost, central, and westernmost portions of the dike. X-ray analyses by the Virginia Division of Mineral Resources noted the



Figure 10. Site of mica peridotite, located west of Front Royal, Warren County.

following: ankerite, calcite, chlorite, clays, garnet (andradite), magnetite, phlogopite, pyrite, quartz, rutile, serpentine, and talc. Ten soil samples along two traverses were analyzed by atomic absorption and indicated significant peaks for chromium (300 to 400 ppm) and nickel (300 to 550 ppm) in both traverses (Rader and Biggs, 1976).

Southworth and others (1993) have described the dike as a mica pyroxenite dike similar to dated late Jurassic dikes in the Appalachians and they indicated the dike to be a kimberlite. At the present time, only scattered pieces of weathered dike rock are present on the site, which extends across properties owned by half a dozen different landowners.

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Permitting industrial mineral operations in the West

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Abstract

Operators permitting a new mine or expanding an existing mine in the western United States must become involved with a myriad of Federal, state, and local agencies. Their effort to obtain the necessary permits is driven by Federal, state, and local regulations; this labor intensive process can be facilitated with pre-project planning. The purpose of this report is to assist in such planning. This report describes the major permitting requirements for mining operations in the western United States, focusing on Wyoming; emphasizes the need to lay out a realistic permitting schedule; and describes a procedure, called an opportunities analysis, that may be used to facilitate the permitting process, minimize costs, and maintain the project schedule.

First we provide an overview of permitting requirements, including Federal requirements for an environmental impact statement/environmental assessment, the state permit to mine, threatened and endangered species and cultural resources clearance, air quality permits, water quality permits, and miscellaneous Federal, state, and county approvals or permits. Second, we present a generalized permitting time line–permitting a new mine in Wyoming may take up to three years, depending on the complexity of the proposed mine and the environmental issues associated with project development. Finally, we present an overview of an opportunities and constraints analysis, a procedure that enables mine developers to identify and troubleshoot environmental issues early in the permitting process and minimizes the potential for costly delays. This analysis includes identification of reasonable development scenarios; consultations with agencies, groups, or individuals with a stake in the project; environmental data collection and analysis; and selection of a preferred development scenario.

Introduction

Permitting a new mine or expanding an existing mine in the western United States requires that operators become involved with a myriad of Federal, state, and local agencies. The level of effort needed to obtain the necessary permits is driven by Federal, state, and local regulations, and, while the process is labor intensive, it can be facilitated with pre-project planning. To assist in such planning, the purposes of this report are:

1) to describe the major permitting requirements for mining operations in the western United States, with a focus on Wyoming;

2) to emphasize the need to lay out a realistic permitting schedule; and

3) to describe a procedure, called an opportunities and constraints analysis, that may be used to facilitate the permitting process, minimize costs, and maintain the project schedule.

Because each mine has somewhat different permitting requirements, the framework presented herein is not necessarily comprehensive, but provides general guidance for achieving compliance with Federal, state, and local environmental laws.

The scope and complexity of the permitting process is dictated by the complexity of the proposed project, characteristics of the proposed development site (including land ownership, mineral ownership, and local environmental features), applicable regulations, and the attitudes of principal stakeholders. For the purposes of this report, we have assumed the following scenario: the proposed development would be a new mine on a mixture of Federal, state, and private land,

and the mineral resource is owned by the Federal government. Facilities required for the new mine would include quarries, a mill or processing plant, utilities (electricity, gas, water, sewer), roads, an office complex (including offices, changing rooms, a lunch room, sanitary facilities, etc.), a warehouse, maintenance shops, fuel oil storage and dispensing facilities, and miscellaneous storage areas.

We have also assumed that there are no known critical attributes associated with the site (i.e., exceptionally valued natural or cultural resources, unique environmental features, or issues of major public concern). Disturbance associated with the proposed new mine would be greater than five acres. As a rule of thumb, however, public scrutiny of a proposed project generally increases with increasing project size and tends to extend the permitting process to allow for greater public input.

Overview of permitting requirements

A summary of authorizing agencies and actions is presented in **Table 1**. Major permitting requirements are discussed below.

Environmental assessment/ environmental impact statement

In the scenario presented above, the proposed operation would be, at least in part, on Federal land managed by a Federal agency (e.g., U.S. Bureau of Land Management, U.S. Forest Service). Under the Federal Land Policy and Management Act of 1976, the Federal land management agency would be required to approve a Plan of Operations or a Plan of Development before the new mine could be developed. Approval of these plans by the agency constitutes a major Federal action and is thus subject to compliance with the National Environmental Policy Act (NEPA) (Public Law 91-190, 42 United States Code 4321 et seq., as amended) and Council of Environmental Quality Guidelines for enforcing the National Environmental Policy Act (40 Code of Federal Regulations Parts 1500-1508), which require the agency to prepare an environmental assessment (EA) or an environmental impact statement (EIS) prior to approval of any major Federal action. Even if the proposed operation was 100% on private land, access roads, transmission lines, and other facilities may cross Federal land and require a right-of-way grant or a temporary use permit, thus, activating Federal involvement. Similarly, proposals to mine Federally owned mineral deposits or projects utilizing Federal funding also prompt Federal involvement and trigger the need for compliance.

The primary purpose of an EA or EIS is to identify the expected environmental consequences of a major Federal action so the managing agency may make an informed decision, fully considering the potential environmental consequences of the proposed action. An EA is prepared when the significance of impacts from the proposed action is unknown, but potentially significant. If the EA does not uncover any potentially significant impacts, the agency writes a Finding of No Significant Impact and may use this as a partial basis for approving the project. If potentially significant impacts are identified, or the significance of potential impacts remains uncertain following an EA, the agency must prepare an EIS. For example, a proposal to develop a quarry adjacent to a wild and scenic river would require the managing agency to critically evaluate the potential for soil erosion and sedimentation in the river, possible impacts to fisheries and water quality, effects on recreational users, etc. While these impacts may be considered significant, the agency is not prohibited from approving the project, but must use the EIS as a tool for making an informed decision.

The scenario in which, after EA preparation, the lead agency determines that an EIS is necessary is costly and time consuming. If there is a high likelihood of significant impacts, time and money can be saved if the lead agency skips the preparation of an EA and evaluates potential project impacts using an EIS. Typically, the lead agency can make an early determination as to whether an EA will be adequate or an EIS will be required, such that preparation of both an EA and an EIS would rarely occur.

The Federal land management agency administering the land or minerals to be mined becomes the lead agency for the NEPA process and is ultimately responsible for the decision to approve or deny the project. Other Federal agencies may be also be involved; for example, the Bureau of Indian Affairs may have a stake in the proposed action and would participate in the NEPA process as a cooperating agency. Upon initiation of the NEPA process, numerous other agencies (e.g., the state game and fish department) automatically become involved. The Federal agency must seek input from cooperating and other agencies during the environmental review process, to ensure that all relevant data are included in the NEPA document.

Table 1. Federal, state, and county permitting actions and regulatory authority.

Agency	Action	Authority[1]
Federal Land Management Agency	Prepare environmental assessment/environmental impact statement.	National Environmental Policy Act (42 USC 4321 et seq., 40 CFR Parts 1500-1508, 10 CFR 1021) Executive Order 11514 as amended by Executive Order 1191; Federal Land Policy and Management Act of 1976, as amended, Public Law 94-579.
	Issue right-of-way grant.	United States Department of Interior/Department of Agriculture/Department of Transportation Public Law 96-487 Federal Register Notice 6-3-81.
	Issue license to explore.	Leasing of Solid Minerals other than coal and oil shale (43 CFR 3500).
United States Environmental Protection Agency	Oversee National Environmental Policy Act and all permitting processes.	See authorities for other federal agencies.
	Permit treatment, storage, or disposal of hazardous wastes.	Resource Conservation and Recovery Act.
United States Fish and Wildlife Service	Review impact on federally listed or proposed threatened and endangered species of fish, wildlife, plants, and migratory birds.	Fish and Wildlife Coordination Act of 1934, as amended 1946, 1958, 1977 (16 USC 661-667e); Endangered Species Act of 1973 (16 USC Sections 1531 et seq.); Migratory Bird Treaty Act of 1918, as amended (16 USC 703 et seq.); Eagle Act (16 USC 668-668d), Executive Order 11990.
United States Army Corps of Engineers	Issue Section 404 Permit for placement of dredged or fill material into waters of the United States.	Section 404, Clean Water Act of 1977, amended 1987 (33 USC Sections 1251-1376).
Wyoming Game and Fish Department	Review impact on threatened and endangered species, wildlife, and wildlife habitat.	Fish and Wildlife Coordination Act of 1934, as amended 1946, 1958, 1977 (USC 661-667e).
Wyoming Department of Environmental Quality		
Air Quality Division	Issue permit to construct and modify.	Clean Air Act (42 USC 7609).
Industrial Siting Council	Issue industrial siting permit.	Industrial Siting Act W.S. 35-12-101 through 119.
Land Quality Division	Isssue permit to mine.	W.S. 35-11-112 (a)(i); W.S. 16-4-101 through 115.
	Issue permit to explore.	W.S. 35-413 through 414, W.S. 35-11-404.
Water Quality Division	Issue Section 401 certification for stream crossings.	Clean Water Act of 1977, amended 1987 (33 USC Sections 1251-1376).
	Issue stormwater discharge permit.	Clean Water Act of 1977, amended 1987 (33 USC Sections 1251-1376); Wyoming Water Quality Rules and Regulations Chapter XVIII.
	Notification of accidental release of hazardous substances into waters of the state.	W.S. 35-11-301 and 35-11-302.
Wyoming State Historic Preservation Office	Consult with federal agency on site eligibility for National Register of Historic Places and the effects of the project on eligible sites.	National Historic Preservation Act of 1966, as amended (16 USC 470); Historic Sites, Buildings and Antiquities Act of 1935, as amended (16 USC 461-467); American Indian Religious Freedom Act (42 USC 1996); Executive Order 11593; Antiquities Act of 1906 (16 USC 431-433, 43 CFR 3); Archaeological Resources Protection Act (16 USC 470aa et seq., 43 CFR 7); Archaeological and Historic Data Preservation Act of 1974 (16 USC 469-469c); Native American Graves Protection and Repatriation Act of October 1990 (25 USC 3001-3013).
County	Issue special use permit.	County rules and regulations.
	Issue building permit.	County rules and regulations.
Department of Transportation	Issue permit for oversized/overweight haul trucks.	State rules and regulations.

[1] WS=Wyoming Statute
USC=United States Code
CFR=Code of Federal Regulations

While it is an agency's responsibility to prepare the EA or EIS, it is often infeasible for them produce the document in time to meet industry schedules. To maintain a reasonable project schedule, a mining company may enlist the services of an environmental consultant (with approval from the agency) to prepare the EA or EIS. Acting as an independent third party, the contractor uses a team of specialists, each of which has a counterpart within the agency, to objectively evaluate the environmental consequences of a proposed action and write the NEPA document for the agency. Ultimately, the agency must accept the document as its own. Once the agency specialists have reviewed and accepted the EA/EIS, it becomes the agency's document and is made available for public review.

State permit to mine

Rules and regulations concerning the permitting of mines other than coal mines are under state authority, so long as state programs comply with Federal laws (e.g., the Clean Air Act, the Clean Water Act). Because there are no Federal laws uniquely governing non-coal mining, state programs (and level of detail required for a state permit to mine) vary considerably from state to state. The Surface Mining Control and Reclamation Act of 1977 (30 United States Code 1201 et seq.) specifically regulates coal mining, and some states, including Wyoming, have modeled non-coal programs after this Act.

In Wyoming, the state agency responsible for regulating all mining activities is the Wyoming Department of Environmental Quality, Land Quality Division. The information required for a mine permit application to the Wyoming Department of Environmental Quality includes:

an adjudication file containing landowner agreements;

a discussion of existing land uses in the project area;

complete historical, archaeological, and paleontological evaluations;

a description of the area's climate;

information pertaining to topography, geology, and hydrology;

a baseline soil survey, including laboratory analyses to determine soil suitability for reclamation;

a baseline vegetation survey, which may include quantitative sampling and, thus, must be completed during summer;

baseline wildlife surveys, which may have seasonal restrictions and, thus, could affect the overall schedule;

wetland delineations; and

mining and reclamation plans.

The permit application also includes an analysis of potential impacts of mining and reclamation on the environment and demonstrates the permittee's ability to meet state performance standards during and after mining, and the permittee's commitment achieving postmining land use and management goals.

In Wyoming, the state permit to mine typically requires more data collection and analysis than other permits or compliance documents for the same development; therefore, data collected for the permit to mine can be used to meet other requirements. For example, the permit application requires baseline information and impact assessment for many of the resources analyzed in the EA or EIS for the Federal agency; thus, scheduling the concurrent preparation of the state permit with other permits or documents may facilitate the permitting process.

Threatened and endangered species clearance

Threatened and endangered species clearance must be obtained prior to any environmental disturbance related to a Federal action, pursuant to the Endangered Species Act of 1973 (16 United States Code Section 1531 et seq.). If any threatened or endangered species occur in the vicinity of the proposed project, a biological assessment must be prepared by the Federal land management agency (or its representative) to assess possible project impacts on these species. If there is potential to adversely affect listed species, the agency must initiate formal consultation with the U.S. Fish and Wildlife Service (FWS). Consultation is necessary to ensure that the Federal agency does not jeopardize the continued existence of a listed or proposed threatened or endangered species or adversely modify or destroy designated critical habitat of such species. Based on the biological assessment, the FWS issues a biological opinion that the proposed project is or is not likely to jeopardize the continued existence of the listed species. The Federal land management agency must consider the results of consultation and the biological opinion in its decision to approve or deny the proposed action.

The biological assessment can be prepared simultaneously with the NEPA document, since both documents contain similar information concerning threat-

ened and endangered species, as long as the schedule for the EA/EIS allows for the collection of necessary seasonal data.

Cultural resources clearance

To comply with Section 106 of the National Historic Preservation Act of 1966, as amended (16 United States Code 470), cultural resources clearance must be obtained in all proposed disturbance areas. A Class I survey (a database and literature search) may be adequate for the EA or EIS; however Class III field inventories are required prior to any surface disturbance. The cost and time it takes to obtain cultural resources clearance depends on the number of sites found within the project area, the importance or significance of sites that may be disturbed, and whether or not the sites can be avoided during the course of the project. Because the cultural work is very field-oriented and dependent on the number of sites within a given area, allowing a degree of schedule and budget flexibility for unexpected contingencies is essential.

Air quality

Under the 1990 Clean Air Act Amendments, state programs must meet the minimum Federal air quality standards, but are not precluded from adopting standards that are more stringent. States are given some leeway to adopt permitting and enforcement rules and regulations; therefore, the laws and interpretations used to apply and enforce the 1990 Clean Air Act Amendments vary somewhat from state to state.

In Wyoming, the Department of Environmental Quality, Air Quality Division, issues air quality permits for mine development or expansion. A Permit to Construct and Modify is required for construction of a new facility or emission source or modification of an existing facility or emission source. The permit application requires specific information concerning facilities planned for construction, when and where the facilities would be constructed, and the type and amount of air quality emissions that would be expected. The permit application also specifies the type of equipment to be used, the duration of use, and what types of pollution control equipment or technologies (e.g., water spray, paving, chemical suppressants) would be used to control emissions.

Once the permit has been issued, an operator can construct and operate as long as the conditions outlined in the permit are met and state air quality standards are maintained. In some cases, an operator may be required to implement an air quality monitoring program to assure compliance with permit conditions and state standards.

Water quality

Water quality protection is typically administered by states under the authority of the Clean Water Act of 1977, as amended (33 United States Code Sections 1251-1376). The Wyoming Department of Environmental Quality is responsible for:

approval of sewage treatment facilities;

compliance with Section 401 of the Clean Water Act which requires certification for stream crossings;

dam or diversion construction permits;

stormwater pollution prevention plans (which allow the operation to be covered under the state's National Pollution Discharge Elimination System permit);

approval of mine water disposal; and

enforcement of state water quality rules and regulations for surface and groundwater.

Other permits/agencies

The United States Environmental Protection Agency oversees permits under the Resource Conservation and Recovery Act for treatment, storage, or disposal of hazardous waste. In underground mines where radon may be a concern, the Environmental Protection Agency requires monitoring and reporting. The Agency also oversees the NEPA process.

Wetlands and waters of the United States are regulated under the Clean Water Act of 1977, amended 1987 (United States Code Sections 1251-1376), the Rivers and Harbors Act, and Executive Order 11990, which mandates no net loss of wetlands in the United States. If wetlands occur in areas to be affected, wetland delineations must be completed prior to surface disturbance. If any jurisdictional wetlands or waters of the United States would be disturbed by the development, a permit from the U.S. Army Corps of Engineers is required. Mitigation for loss of wetlands (i.e., wetland enhancement or construction) may be a stipulation in the permit.

In Wyoming, if construction costs exceed a specified amount, the facility qualifies as an industrial facility and requires a permit from the Wyoming Department of Environmental Quality, Industrial Siting Council. The application includes much of the same information contained in the NEPA document, with additional information on the social and economic impacts of the proposed facility. As mitigation for adverse economic impacts on local communities, a schedule of compen-

satory payments is calculated and included in the application.

Operators also should expect to meet the permitting requirements of:

Wyoming Office of State Lands and Investments,

Wyoming Department of Transportation,

Wyoming State Engineer's Office, and

county rules and regulations.

Exploration

Both state and Federal agencies require a license or permit to explore. Because exploration involves surface disturbance, many of the issues discussed above for mine development also apply to exploration. The Federal land management agency is required to prepare an EA and/or EIS; cultural resources and threatened and endangered species clearance is required; exploration and reclamation plans must be developed, etc. However, since disturbance associated with exploration is typically minor, the permitting process is generally fairly brief and relatively inexpensive.

Permitting schedule

The need for a mineral product of a certain quality by a specific date is a primary factor driving the development of a new mine or expansion of an existing mine. The permitting schedule can be derived by working backwards from specific target dates:

Identify and schedule the desired objectives (e.g., the plant on-line date, a contract delivery date, etc.).

Identify permitting requirements–determine which permit takes the longest to obtain and what studies or approvals may be prerequisites for permit applications. For example, in many cases the Federal land management agency will not approve a project until the FWS has evaluated potential impacts to threatened or endangered species.

Schedule each major milestone in the permitting sequence.

Allow for unexpected contingencies.

Figure 1 is a generalized time line for permitting a new mine. Submission of the Plan of Operations or a Right-of-Way Grant application triggers the start of the Federal NEPA process. Preparation of the EA or EIS and the state permit can commence very early in the development process, even before exploration has been completed. Generally, an EA takes between six months and a year to prepare; project approval takes about 14 months if there are no significant impacts and no major controversies. EIS preparation, review, and decision generally take between one and three years. If significant impacts are likely, it may be prudent to short-cut the NEPA process by directly preparing an EIS instead of an EA and then an EIS.

Because the Wyoming state permit to mine requires detailed baseline information, it generally takes 10 to 20 months to prepare the permit application. Wildlife and vegetation surveys must be completed within specified seasonal periods, so data collection must be planned to accommodate the necessary fieldwork. Once the permit is submitted, the Department of Environmental Quality, Land Quality Division, reviews it for completeness and technical adequacy. It is possible to undergo several rounds of revisions to complete an application, and the agency may take 90-120 days to complete each review. The mine permit typically takes 30 to 36 months to complete. However, other permits can be scheduled within this overall timeframe. Because of the overlap in data requirements among various permits, efficiencies that save time and reduce overall permitting costs can be incorporated into the permitting schedule.

Opportunities and constraints analysis

The costs of environmental compliance may vary widely, and by identifying these costs early on, a company may make an informed decision to select or reject a given development scenario. Unfortunately, it is not possible to foresee all the issues or environmental concerns during planning stages, and unexpected environmental problems may arise. Generally, the agencies work with operators to resolve problems, but, in the meantime, the project is delayed and project expenses increase. This section describes a procedure called an opportunities and constraints analysis that may be used to uncover possible environmental issues during project planning and reduce the potential for costly delays. The goal of an opportunities and constraints analysis is to compare various alternative development scenarios and pick the one that a) meets

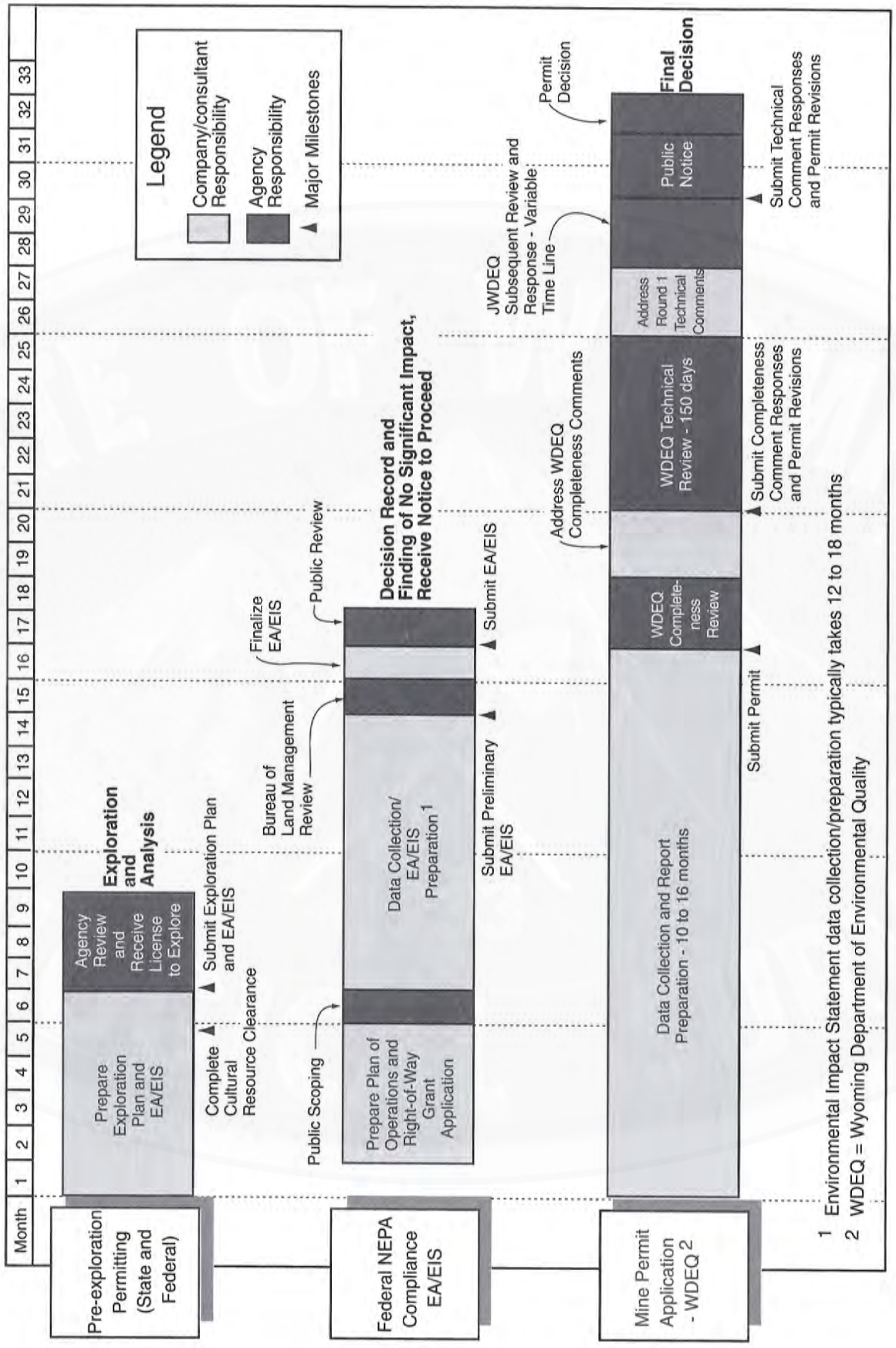


Figure 1. Generalized time line for permitting a non-coal mine.

company objectives, b) provides the greatest number of opportunities (e.g., the presence of an existing right-of-way for the mine's power line), and c) avoids constraints (e.g., a large number of stream crossings). By identifying possible fatal flaws, maximizing opportunities, and avoiding or compensating for constraints, overall schedule and project development costs are minimized.

An opportunities and constraints analysis involves four steps (which are explained in detail, below): identification of reasonable development scenarios; consultation with agencies, groups, or individuals with a stake in the project; environmental data collection, including identification of data gaps and collection of baseline data (if necessary); and analysis of environmental data and selection of a preferred development scenario.

The opportunities and constraints analysis is performed by the developer prior to initiating formal permitting. Once complete, the agencies may be presented with a proposal that has been screened for environmental problems, thereby expediting project start-up and reducing the potential for unexpected obstacles.

Identifying reasonable development scenarios

Identification of reasonable development scenarios (from a company perspective) is based on an evaluation of mineral quality, quantity, accessibility, proximity to markets, land ownership, etc. Exploration usually involves a feasibility study of several mineral bodies and selection of the deposit that would best meet company goals. The opportunities and constraints analysis introduces another set of variables–environmental considerations–which are factored into the feasibility analysis. Having several promising alternatives is necessary for an effective opportunities and constraints analysis. The types of scenarios analyzed may include various mine or facilities locations, routes for access roads or power lines, and mining methods–i.e., whatever may be appropriate for the specific situation. If several options are feasible from a development perspective, the opportunities and constraints analysis can provide a means to identify the proposed action with the least number of environmental obstacles, thereby reducing the likelihood of disapprovals, delays, and associated project costs. In cases where a particular scenario is so outstanding that no others are considered reasonable, the opportunities and constraints analysis can be used to search for fatal flaws which could delay or stop the project or add significantly to its costs.

The identification of various development scenarios, a process that is performed by the company, should not be confused with the analysis of *alternatives* performed by the lead agency during EA/EIS preparation. As part of the NEPA process, the lead agency selects a set of alternatives, in addition to the proposed action, to analyze. The agency-selected alternatives may or may not correspond to the various scenarios the company evaluated during the opportunities and constraints analysis because they serve a very different function–to assist the agency in making an informed decision concerning a proposed action. Step 1 of the opportunities and constraints analysis is not part of the NEPA process.

Consultation

The development scenarios should be discussed informally with the primary regulating agencies, special interest groups, and appropriate individuals to determine the types of environmental issues and the level of concern that may be associated with each scenario. The objectives of these informal consultations are to identify obvious problems with the various scenarios, to determine possible mitigation requirements, and to gain an awareness of public opposition/support for the project. While many impacts can be mitigated, the relative costs of mitigation or delays caused by public resistance may make one scenario more attractive than another. For example, the Federal land management agency may have information concerning important cultural resources or critical wildlife habitat in the vicinity of a proposed alternative, thereby forewarning the company that extensive mitigation may be required if the project goes forward in this area. All other things being equal, another site may be more cost-effective.

For the opportunities and constraints analysis, consultation is informal and focused on data-gathering and information-sharing, rather than permitting, although the long-term goal–permitting the development–would be the same. Once the parties involved have good information concerning the proposed development and the scenarios being evaluated, they can determine what, if any, site-specific data are available, and the next phase of the opportunities and constraints analysis is well underway.

Environmental data collection

Probably the best way to compile and synthesize existing data for the opportunities and constraints analysis is to develop a matrix of information for each resource under each alternative (**Table 2**). In the ex-

Table 2. Sample evaluation of opportunities and constraints.

Resource	Development Scenarios		
	Mine Site No. 1	Mine Site No. 2	Mine Site No. 3
Geology			
Geologic hazards (number of landslide areas)	4	5	0
Conflicts with other minerals (number of leases)	1	2	3
Soils			
Highly erosive soils (acres)	15	10	5
Saline soils (acres)	5	2	7
Water Quality			
Number of stream crossings	2	3	3
Impacts of mining on subsurface water quality (arbitrary rank)	3	2	1
Land Use			
Number of residences	0	1	2
Acres of wilderness	0	0	25
Available rights-of-way (miles)	15	10	0
Wildlife			
Impacts on threatened and endangered species (acres of potential habitat)	10	10	5
Big game crucial winter range (acres)	25	30	0
Number of raptor nests	8	15	17
Number of sage grouse leks	2	3	1
Vegetation			
Impacts on threatened and endangered species (acres of potential habitat)	20	20	20
Wetlands (acres)	2	5	4
Cultural and Historic Resources			
Number of sites	5	2	2
Visual Resources			
Miles of road from which the site would be visible	15	14	20
Air Quality			
Distance from Class I areas (miles)	50	50	20
Noise			
Distance from nearest residence (miles)	10	2	8
Public Safety			
Extent of measures needed to ensure public safety (arbitrary rank)	2	1	3

ample shown on **Table 2**, impacts to 10 resources were compared at three hypothetical mine sites. Data describing the physical and biological characteristics of each area are collected from existing sources (e.g., U.S. Geological Survey, the Natural Resources Conservation Service, the primary regulating agencies, and published literature). While much of the information needed to evaluate development scenarios is available from existing sources, existing data may be inaccurate or incomplete, and data gaps should be identified at this phase of the analysis. It may be prudent to complete a field reconnaissance or baseline studies as part of the opportunities and constraints analysis, especially if there are issues of special concern; otherwise a problem might not be identified until fieldwork or other in-depth analysis is underway, where it may cause costly delays. For example, the agency may want to evaluate impacts on nesting mountain plovers but may not have nesting data for each mine site. A Class I file search for cultural resources is an excellent resource for determining the potential for archaeological finds, but it is not possible to know the actual number of sites in a given area until the fieldwork is complete. By identifying data gaps and completing baseline work early in the process, the project proponent is well informed about the environmental characteristics of possible development sites and, thus, is in an excellent position to develop a solid proposal that will withstand regulatory and public scrutiny.

Environmental data analysis and selection of a preferred development scenario

Synthesis and analysis of environmental data enables an operator to rank alternatives based on:

1) anticipated costs of mitigation, 2) costs associated with obstacles that would necessitate extensive environmental study, and 3) costs associated with obstacles that would cause permitting delays.

Once the baseline data have been collected and summarized, impacts of mining, road construction, crushing facilities, groundwater withdrawal, and other activities associated with the proposed development on the existing physical and biological resources can be evaluated based on an appropriate quantitative measure. Quantification of impacts is based on site-specific circumstances. For example, wildlife impacts may be measured as the amount of habitat lost under each alternative, the number of raptor nests within 0.5 mile of the project area, the miles of trout stream affected, etc. Noise impacts may be assessed based on the number of residences near each alternative. Opportunities (such as the possible use of existing rights-of-way for roads and power lines) and beneficial impacts (such as increased employment) also are identified.

Typically, a qualitative or semi-quantitative measure is used to assess the possible costs associated with impacts to natural or cultural resources. For example, in **Table 2**, Mine Site 1 has 15 miles of right-of-way available for road or power line construction, compared with 10 and 0 miles for Mine Sites 2 and 3, respectively. Therefore, Site 1 would cause the least amount of new right-of-way disturbance, and costs associated with environmental compliance (i.e., cultural resources clearance) would be lower. If the FWS requires raptor nest relocation, Site 1 would be least expensive. If public opinion strongly opposes mine development near residences, Sites 1 and 2 would be environmentally preferred.

Selection of the preferred scenario is based on information gathered during exploration and the subsequent environmental analysis. By keeping a running total of costs and maintaining an awareness of possible obstacles and public perceptions (which may determine the relative importance of various issues to agency decision makers), it may be possible to identify a path of least resistance for acquiring the necessary permits in a timely and cost-effective manner. The opportunities and constraints analysis is quite flexible–it can be tailored to provide very general or very detailed information, depending on project needs.

Upon identification of a preferred scenario or scenarios, permit applications may be submitted to the appropriate agencies, and formal environmental reviews may begin. The opportunities and constraints analysis provides permit/document preparers with sufficient data to propose a course of action with enhanced probability of timely approval.

Summary

The permitting framework presented herein can be summarized as follows:

1) Identify permitting needs.

2) Develop a permitting time line.

3) Identify reasonable development scenarios.

4) Complete an opportunities and constraints analysis with agency and public input. Importantly, identify "red flags" associated with inadequate environmental

data and/or agency/public concerns–collect the necessary data and address issues early in the opportunities and constraints process.

5) Rank development scenarios considering environmental costs, and select a preferred development scenario.

6) Begin formal review and permitting of the scenario with the highest likelihood of approval, minimizing delays and costs associated with the permitting process.

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GEOLOGICAL SURVEY

How specifications govern the availability of natural aggregate

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Abstract

During 1994 the United States consumed over 2.3 billion tons of natural aggregate (crushed stone and sand and gravel). During the next decade, the restoration and rehabilitation of an aging infrastructure, as well as new construction, will require enormous amounts of aggregate, much of which must meet strict specifications for new high-performance construction materials. If aggregate quality is unspecified, nearly all surface material could be considered as a source. However, most construction projects require aggregate that meets certain specifications. Those specifications can significantly affect the availability of aggregate.

The aggregate assessment of the Mimbres Resource Area (MRA), southwestern New Mexico, demonstrates how severely specifications can limit aggregate availability. The MRA lies within the Basin and Range Province, an area that is blessed with abundant supplies of natural aggregate. As part of the assessment process, the MRA was evaluated to identify areas underlain by aggregate that meet requirements for use in Portland-cement concrete. Only about 10% of the MRA is underlain by aggregate that is likely to be suitable for use with Portland-cement concrete.

The Mimbres assessment used specifications that exist today. However, future demand most likely will require even higher-quality aggregate. For example, although today's concrete has a typical compressive strength in the range of 3000 to 5000 pounds per square inch (psi), it is reasonable to expect future high-performance concrete to have compressive strengths of 15,000 psi. The purpose of aggregate in concrete is to take up space; thus, the strength limits of aggregate need to be suitably matched with increasing strength limits of cement concrete. A rule-of-thumb is that aggregate should be two to three times as strong as the final concrete product. The anticipated compressive strength of high-performance concrete will have strength values that approach or exceed the compressive strength of many of the rocks currently used as aggregate. New strength requirements could limit the availability of aggregate.

*Recently the Strategic Highway Research Program (SHRP) completed a new system for specifying asphalt materials which is referred to as Superpave (**Su**perior **Pe**rforming Asphalt **Pave**ments). It was the consensus of the SHRP pavement experts that certain aggregate characteristics were critical for producing high-quality hot mix asphalt. In order to reduce pavement deformation (rutting), one of the specifications of Superpave limits the amount of round-grained natural sands that can be used by specifying that the fine aggregate contain a specified amount of angularity. Preliminary observations are that cubical particles in manufactured sand resulting from the processing of crushed stone (100% fractured faces) from many deposits of high-calcium limestone will not meet some Superpave specifications. Many of these deposits have long histories of providing suitable aggregate for use in interstate-type asphalt pavements. Although more information and research is needed, current Superpave requirements may drastically reduce aggregate availability in certain geographic areas.*

Introduction

Sand and gravel is produced in every state in the United States, and crushed stone is currently produced in every state except two (Delaware and North Dakota) (Tepordei, 1996). During 1994 the United States consumed about 1.0 billion short tons of sand and gravel and about 1.3 billion short tons of crushed stone, for a total of about 2.3 billion tons of natural aggregate (Tepordei, 1995). During the next decade, the restoration and rehabilitation of an aging infrastructure, as well as new construction, will require enormous amounts of aggregate.

A common perception exists that natural aggregate of an appropriate quality is abundantly available, at a reasonable cost, wherever it is needed. This may be true if one considers that aggregate quality is unspecified. Nearly all surface materials, even ones such as ice and rock salt, can be considered a source of construction materials. Ice has been used to build igloos and ice castles; and rock salt has been used as an aggregate mixed with cement and supersaline water to plug leaks in excavations in salt domes. However, almost all uses of aggregate require that the aggregate meet minimum quality specifications. During 1994, 83% of aggregate produced in the U.S. was used in Portland-cement concrete, bituminous mixes, and other construction projects that require specific quality parameters (U.S. Bureau of Mines, 1995).

Aggregate quality

Natural aggregate varies widely in quality, depending on the source. Suitable aggregate should have strength properties that resist mechanical breakdown resulting from the action of mixers, mechanical equipment, and (or) traffic. Aggregate should be physically sound and be able to withstand stresses caused by repeated freezing and thawing or wetting and drying. These properties are generally controlled by the individual mineral constituents of the rock particles, the strength with which these minerals are bound together, and the abundance of fractures.

Aggregate should not change chemically once in place. However, some aggregates contain minerals that chemically react with or otherwise affect the Portland-cement concrete or bituminous concrete. One of the most destructive reactions in Portland-cement concrete is the alkali-silica reaction (ASR). The ASR is a reaction between specific types of silica in the aggregate and alkali in the cement pore solution. Under specific conditions, certain types of silica readily react with the alkali hydroxide solutions in the cement pastes, producing alkali-silica gel. This gel may imbibe water and swell. The swelling causes expansion of the aggregate and subsequent expansion and cracking of the concrete which can dramatically disrupt the surface of the concrete; in extreme cases expansion can cause complete destruction of structures (Dolar-Mantuani, 1983). Damage due to expansion may be visible after several months, but with some types of rocks expansion may occur after 30 years or more.

In asphaltic concrete the electrochemical properties of rock relate to the bitumen's ability to adhere to the aggregate. Many investigators agree that siliceous materials acquire a negative charge in water (Hoiberg, 1965). In contrast, some types of aggregate such as limestone and other calcareous materials have a positive surface charge in the presence of water. Some rocks contain both silica (which acquires a negative charge) and calcium, magnesium, aluminum, or iron (which acquire a positive charge). Charged molecules and atoms in the water lower the interfacial tension and consequently lower the adhesion tension (Smith and Collis, 1993). Such rocks include trap rock, basalts, porphyries, and siliceous limestones (Hoiberg, 1965).

Any rock that normally would be considered adequate for most uses can be rendered inadequate if it has been weathered. Even slightly weathered rock may be somewhat weaker than in its fresh condition, and its aggregate properties may be significantly degraded (Fookes, 1980).

The properties that one looks for depend primarily on the uses to which the aggregate will be put. Departures from any quality requirements can render aggregate unsuitable for specific uses.

Mimbres Resource Area

The U.S. Geological Survey, in cooperation with the New Mexico Bureau of Mines and Mineral Resources, conducted an assessment of mineral resources in the Mimbres Resource Area (MRA), southwestern New Mexico. The project included an assessment of natural aggregate resources (sand and gravel

and crushable stone). The MRA is in the Basin and Range Province, an area blessed with abundant supplies of natural aggregate. An objective method was designed to assess the area utilizing available geologic maps in digital format, without the benefit of supporting field work (Langer and others, 1996). The evaluation included estimates of the probable quality of bedrock lithologies and unconsolidated deposits for use as aggregate for Portland-cement concrete. Specifically, those deposits that were likely to contain coarse clasts that were either physically weak or contained minerals that may participate in the alkali-silica reaction were identified as poor-quality aggregate.

About 36% of the MRA is underlain by exposed or near-surface bedrock. When the quality of the bedrock was considered, only about 4.4% of the MRA remained as potential sources of high-quality crushed-stone aggregate. About 64% of the MRA is underlain with unconsolidated materials, but only about 23% of the MRA is underlain with unconsolidated materials that contain significant amounts of gravel. When the quality of the gravel clasts was considered, only about 6% of the MRA contained abundant potential sources of high-quality sand and gravel.

Combining the figures for crushed stone and sand and gravel, only about 10% of the MRA is underlain by aggregate that is likely to be suitable for use in Portland-cement concrete. Because of the type of data used to assess the aggregate potential of the MRA and the lack of accompanying field studies, the results of the assessment may only be general indications of the aggregate availability in the MRA. Areas designated as lower quality may actually contain some higher-quality aggregate, or lower-quality aggregate could be processed to meet higher quality requirements. Such areas may contain aggregate that could be used for purposes that have less stringent quality requirements. However, the resource maps prepared for the assessment demonstrate the dramatic affect quality can have on aggregate availability.

High-performance construction materials

As we progress into the 21st century, demand will be accompanied by specifications requiring higher-quality aggregate. Although today's concrete has a typical compressive strength in the range of 3000 to 5000 pounds per square inch (psi), high-performance concrete may have compressive strengths exceeding 15,000 psi. Such strength parameters could lead to substantial savings in highway and bridge construction and other construction projects. For example, the Federal Highway Administration concluded that using 10,000 psi concrete could reduce the number of supporting girders for a 150-foot bridge from nine to four. Furthermore, increasing the durability of America's roads and bridges by just 1% could lead to long-term savings as great as $10 billion to $30 billion over a 20-year period (Civil Engineering Research Foundation, 1993).

The purpose of aggregate in cement concrete is to take up space; thus, the strength limits of aggregate need to be suitably matched to increasing strength limits of concrete. The anticipated compressive strength of high-performance concrete approaches or exceeds the compressive strength of many of the rocks currently used as aggregate. Examples of typical compressive strengths for fresh, unweathered rocks commonly used as aggregate are: limestone, 2000 to 37,000 psi; sandstone, 5000 to 36,000; granite, 14,000 to 45,000; and basalt, 16,000 to 49,000 (Winkler, 1973). A good average aggregate has a strength several times greater than that of the concrete (Fookes, 1980). Imposing that requirement for high-performance concrete could eliminate large amounts of aggregate from use.

Superpave requirements

An example of how specifications may limit existing permitted aggregate resources is developing today. Recently the Strategic Highway Research Program (SHRP) completed a new system for specifying asphalt materials. The final product of the SHRP asphalt research program is a new system referred to as Superpave (**Super**ior **Per**forming Asphalt **Pave**ments).

While no new aggregate tests were developed, current methods of selecting and specifying aggregates were refined and incorporated into the Superpave mix design system. It was the consensus of the SHRP pavement experts that certain aggregate characteristics were critical for producing high-quality hot mix asphalt. In order to reduce pavement deformation (rutting), Superpave limits the amount of round-grained natural sands that can be used by specifying that fine aggregate (particles smaller than 2.36 mm) meets specified minimum void requirements as tested by

ASTM Standard C 1252-93 (American Society for Testing and Materials, 1994). Following this standard, fine aggregate is allowed to flow through a funnel from a fixed height into a measured cylinder. The uncompacted void content is calculated, which relates to the angularity of the fine aggregate. A large void content indicates aggregate with greater angularity, less sphericity, rougher surface texture, or a combination thereof. A small void content is associated with more rounded, spherical, smooth-surfaced fine aggregate, or a combination thereof. By setting a requirement for a large void content, Superpave specifications can limit the amount of round-grained natural sands in the mix.

Superpave requirements may inadvertently eliminate some presently suitable types of manufactured sand from use in interstate-type asphalt pavements. Preliminary observations are that some sand manufactured by crushing high-calcium limestone deposits do not meet Superpave angularity specifications for high traffic surface mixes. Highly cubical particles of crushed limestone products manufactured by impact-type crushers can give the same C 1252 test results as some round-grained natural sands. Aggregate producers and Departments of Transportation in Georgia, Illinois, Indiana, Iowa, Kentucky, Ohio, and Tennessee have expressed concern that permitted limestone deposits in their states may be eliminated from consideration for use in interstate-type asphalt pavements, even though many of the deposits have long histories of providing suitable aggregate. On the other hand, sand manufactured by dolomite may meet Superpave specifications (Jahn, 1996). (At the time of this writing, only about half the Departments of Transportation in the United States have experience with Superpave aggregate specifications.)

The suitability of large amounts of aggregate may be affected by the Superpave guidelines. About 71% of the crushed stone produced in the United States during 1994 was either limestone and dolomite (U.S. Bureau of Mines, 1995). Commonly manufactured sand is a coproduct with the production of crushed stone. By reducing the marketability of sand from crushed limestone, the economics of an entire crushed stone operation may be affected. This may result in the preferential development of dolomite over limestone. Although some dolomite has strength parameters equivalent to limestone, some varieties of dolomite, especially those where the original dolomite has undergone dedolomitization, can have significantly lower compressive strength values (Williams and McNamara, 1992). In addition, certain dolomitic limestones are susceptible to an alkali-carbonate reaction which is similar to the alkali-silica reaction (Hadley, 1961). Consequently, although more information is needed, Superpave requirements, in their current form, may drastically reduce aggregate availability.

Various geologic phenomena may explain why the impact process of crushing limestone creates less angular sands than crushing dolomite. Limestone is highly susceptible to dissolution by even slightly acidic water; dolomite is much less reactive. Dolomite is chiefly a rock that was formed during diagenesis of limestone, $CaCO_3$, where calcium ions in the lattice are replaced by magnesium ions to form the mineral dolomite, $CaMg(CO_3)_2$. In some cases the original fabric of the limestone is preserved. If dolomitization occurs late in the rock history, the original structure of the rock may be obscured. Dolomites may also be subjected to a process known as dedolomitization, where dolomite is replaced by calcite. Both dolomitization and dedolomitization can result in volume changes and brecciation of the rock (Williams and McNamara, 1992). How any or all of these phenomena affect rock properties is the subject for research.

Effects of failing to meet specifications

If aggregate from existing operations is no longer able to meet new specifications, there are options. The aggregate manufacturing process can be reconfigured to produce more angular products. Existing inferior aggregate can be upgraded or replaced with aggregate from other existing (and most likely more distant) operations. New operations with resources that meet specifications can be opened.

Cubic limestone particles are similar to round particles in that they fit in a very tight configuration; angular particles have corners which cause them to fit in a loose configuration with many voids. The rougher surface of dolomite particles, even when the particles are cubic, allow them to fit into a loose configuration. The closer to a cubic shape crushed stone and fines are, the stronger they tend to be. Consequently, current industry practice commonly favors configuring the manufacturing process to create cubic-shaped clasts. By reconfiguring the processing techniques used to crush limestone, it may be possible to create more angular particles that will meet Superpave specifications. However, making products more angular may downgrade the strength of the particles.

Supplying aggregate from more distant operations can add significantly to the cost of the final product. Costs for transporting aggregate by truck are dependent on local market conditions, but are approximately 25 cents per cubic yard per mile, or 19 cents per ton per mile. A 25-mile haul will double the cost of sand and gravel. A 50-mile haul will double the price of highly-processed aggregates (mixes) and nearly triple the cost of raw sand and gravel (Fowler, 1988). To demonstrate how costs can add up, consider that one lane-mile of a typical interstate highway construction requires nearly 20,000 tons of sand, gravel, and crushed stone. Although construction costs are highly variable and difficult to obtain, figures provided by Tepordei (Barsotti and others, 1995) can be used to calculate that highway construction costs about $1,000,000 per lane mile. Every additional 25 miles that aggregate is hauled will increase the cost of one lane-mile of highway by about 10%.

If aggregate is to be produced from new sources certain conditions must be met. In addition to meeting new quality requirements, the aggregate must be: (1) accessible—aggregate must physically be able to be mined, and must be accessible to transportation systems and to markets; (2) permitable—the site must qualify for all necessary permits and the public must be convinced that the operation can take place without adversely affecting the environment or their lifestyle; and (3) economical—the operation must be profitable considering all costs including acquisition, operation, compliance with regulations, and reclamation (Dryer, 1976; Banino, 1994). These requirements make opening a new operation a complicated process that can cost millions of dollars and take many years.

Summary

On a national scale there are very large amounts of natural aggregate that will meet unspecified uses. However, when aggregate is subjected to specific quality requirements, the availability can be dramatically reduced. Furthermore, new more stringent specifications may eliminate some existing permitted sources of aggregate from use. Therefore, when making resource estimates it is important to consider the quality of the resources, and when considering new aggregate operations it may be beneficial to locate resources that meet anticipated future requirements as well as current requirements.

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Industrial minerals on Federal lands: Locatable, leasable, or salable under Federal law?

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Abstract

Federal lands in the United States administered by the Department of Interior, Bureau of Land Management and Department of Agriculture, Forest Service, total about 461 million acres; 191 million acres are administered by the Forest Service, and 270 million acres are administered by the Bureau of Land Management. The vast majority of these public lands are located in the western USA and Alaska. Although some of these lands are withdrawn from the mining and mineral laws for such uses as wilderness and recreation areas, the majority of them are open for mineral exploration and development.

Several laws dating back to the late 1800s control the disposition of minerals on Federal lands. The 1872 Mining Law set up a system whereby hardrock minerals (gold, copper, lead, etc.) can be staked under mining claims on public lands. Mining claimants have a statutory right to these "locatable" minerals that occur within their claim boundaries. The 1920 Mineral Leasing Act set aside certain minerals including coal, oil, gas, oil shale, phosphate, and sodium and provided a leasing system for their removal from public lands. The 1947 Materials Act and the 1955 Surface Resources Act were later passed to enable the extraction of "salable" minerals that were never covered under the 1872 Mining Law. The 1947 and 1955 Acts removed common varieties of sand, stone, gravel, pumice, pumicite, cinders, clay, and petrified wood from the Mining Law, but provided that certain "uncommon varieties" of these same minerals remain locatable under the 1872 Mining Law. Unfortunately, it has been left up to the legal system to define what makes a mineral "common" versus "uncommon." One particular case heard in the 9th Circuit Court of Appeals, McClarty vs. Secretary of the Interior, *helped set a standard for determining common or uncommon minerals. The basis of the McClarty criteria is whether the mineral in question has a unique property that gives it a distinct and special value, which in turn, is reflected by the premium price that the mineral commands in the marketplace.*

Many industrial minerals fall within the uncommon variety definition, while others are considered common. Minerals such as perlite, gypsum, mica, and garnet, to name a few, are easily categorized as locatable minerals because of their unique characteristics and uses. Others, including certain grades of limestone, silica, clay, pumice, building stone, and decorative stone are much harder to define. Still others, such as potash, phosphate, and sodium, are leasable. Case studies on limestone, pumice, and decorative stone are presented below to illustrate how the common versus uncommon standards have been applied by the courts. In most cases, mineral operators feel it is advantageous to have their deposit on public land classified as a locatable mineral. The reasons for this are several. Disposal of leasable and salable minerals on Federal lands is discretionary on the part of the Federal agency. Locatable mining activities, on the other hand, are non-discretionary, require no Federal royalties be paid, and give claimants the opportunity to patent the Federal mineral and surface estates. Several recent attempts to reform the 1872 Mining Law have been narrowly defeated, but some form of reform legislation may eventually pass into law in the near future.

Introduction–What are Federal lands?

Federal lands owned by the United States government include over 270 million acres (109 million hectares) managed by the U.S. Department of Interior, Bureau of Land Management (BLM), and about 191 million acres (77 million hectares) managed by the U.S. Department of Agriculture, Forest Service (FS). The BLM also manages 570 million acres (231 million hectares) of subsurface mineral estate. The bulk of BLM and national forest lands occur in the western United States and in Alaska (**Figure 1**). The FS has a sizable land base in the eastern United States as well, with over 24 million acres (10 million hectares) located in national forests and grasslands scattered from Texas to New Hampshire. Numerous other Federal agencies control vast tracks of land as well, including the Department of Defense, National Park Service, U.S. Fish and Wildlife Service, Bureau of Reclamation, Army Corps of Engineers, and others.

The majority of Federal lands are open, in one form or another, to mineral entry and/or leasing for exploration and development of the Federal mineral estate. Exceptions to this include "withdrawn" lands within national parks, national wildlife refuges, and many military installations. Depending on the legislative language that established many of these withdrawn areas, some mineral development may take place on these lands as well, especially where valid existing rights are involved (i.e., pre-existing mineral rights or leases for mineral extraction). Sizable portions of BLM and national forest lands have also been withdrawn from the mining and mineral leasing laws for uses such as wilderness, recreational facilities, communication installations, and other similar uses incompatible with mineral development.

Federal land includes several different categories of land status. The bulk of Federal lands are classified as public domain land, which is land that has never left the ownership of the U.S. government. BLM lands are often referred to as public domain or simply, public lands. This category of public land also includes land which has been reserved, or set aside, from the public domain by Congress for certain specific uses such as national forests and national grasslands. Lands that have left the public domain over time through homesteading laws, patenting, exchanges, town site sales, etc., but for one reason or another have been re-

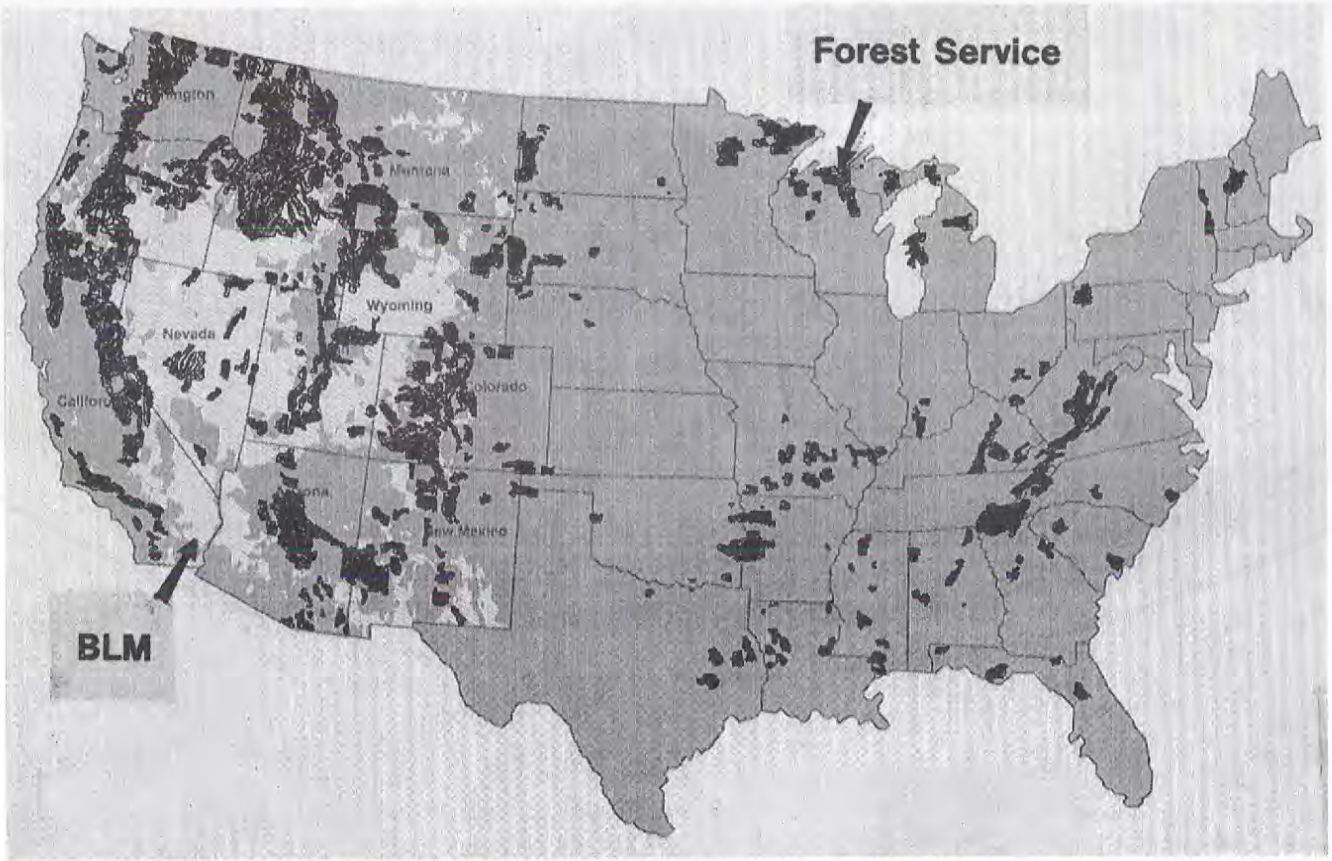


Figure 1. BLM lands (light shaded) and Forest Service lands (dark shaded) in the United States (excluding Alaska). BLM lands total 270 million acres and Forest Service lands total 191 million acres.

acquired by the U.S. government through a land ownership adjustment, are termed "acquired lands." One important distinction of acquired lands is that they are often managed under different mineral extraction laws than public lands. A third category of Federal land includes those lands which are not owned in fee simple by the government. These lands are often referred to as split-estate lands. In some cases, the surface is owned by the U.S. government and the mineral estate is owned by private individuals or companies. In other cases, privately-owned surface lands have Federally-owned mineral estates as part of the deed. This category is often a complex one for the BLM and FS to administer, especially in the eastern U.S. where split estates are more common.

Federal laws affecting industrial minerals

Industrial minerals are a broad category of non-metallic, non-energy minerals. Depending on the specific mineral type and Federal land on which they occur, Federal minerals can fall into one of three legal categories: locatable, leasable, or salable. Locatable minerals are those Federal minerals managed under the 1872 Mining Law, as amended, on public lands. Leasable minerals include energy minerals (other than uranium) and certain non-energy, industrial minerals on all Federal lands, regardless of land status. On acquired Federal lands, hardrock minerals that would be classified as locatable on public lands are considered leasable as well, since the 1872 Mining Law only pertains to public lands. Salable minerals are the third category and include those minerals that are widespread and do not have any particular special characteristics. Examples of salable minerals are sand and gravel and common stone. This paper will examine the differences between these three groups in more detail below.

1872 Mining Law

The 1872 Mining Law was one of the first attempts to organize and facilitate the mining of minerals and disposal of land in the vast American West at the end of the last century. This law made all "valuable" minerals on public lands locatable. Early on, however, problems arose when claimants tried to locate claims for sand, gravel, and common fill. Over time, the Federal courts established that certain minerals were not considered locatable under the law. Locatable minerals under the 1872 Mining Law are generally what we consider the hardrock minerals, which contain elements such as gold, silver, lead, zinc, copper, iron, tungsten, etc. However, many industrial minerals have also been determined to be locatable and come under this law too.

Although it has been amended numerous times since it's inception, the Mining Law still retains many of the fundamental tenants which make it economically attractive to the mining industry and, at the same time, highly controversial. Under the law, mining claims are categorized as lodes, placers, or mill sites, depending on the particular minerals and mining activity involved. Mining claimants who stake a claim have an exclusive property right to the Federal minerals within their claim boundaries. There is no limit to the number of claims an individual or company may hold. Claimants who produce minerals from Federal mining claims pay no royalties on their production. Also, if a mining claim is found to meet certain requirements of the Mining Law, including a discovery of a valuable mineral, the claimant may file for patent, a process whereby the Federal surface and mineral estate is deeded over to the mining claimant and becomes their property.

The BLM is the official record keeper and manager of Federal minerals, and as such, has a leading role in the management of locatable minerals on public lands. Mining claimants are required to submit annual rental fees, or in the case of an exemption for small miners, proof of annual assessment work, for each claim to the BLM. In addition, they are required to file affidavits with the county courthouse in which jurisdiction their claims reside. Both the BLM and FS regulate locatable mineral activities on Federal lands.

1920 Mineral Leasing Act

The 1920 Mineral Leasing Act removed certain minerals from the 1872 Mining Law and established a leasing system for their exploration and extraction. Minerals covered under this law include oil, gas, coal, geothermal energy, oil shale, phosphate, and sodium. Later, other Federal legislation added sulfur in New Mexico and Louisiana, and potash minerals elsewhere to the list of minerals covered under the Leasing Act. These minerals are no longer obtainable through the location of mining claims; rather, Federal leases, permits, or licenses are required for their development. Royalties are also paid to the Federal government on any of these minerals that are produced.

Materials Act of 1947

The Materials Act of 1947 authorized the disposal of minerals by the Federal government that were never intended to be covered under the 1872 Mining Law. These included sand, stone, and common clay. The Act authorized the sale of these materials at fair market values. The Act also authorized the removal of these materials by other government agencies and non-profit organizations, if not used for profit or re-sale.

Surface Resources Act of 1955

In 1955, the Surface Resources Act was passed by Congress. This Act amended the 1947 Act by listing and removing common varieties of sand, stone, gravel, pumice, pumicite, cinders, clay, and vegetative materials from the "valuable" minerals covered under the Mining Law. After 1955, these materials were no longer locatable. The Act however, provided only general guidance to distinguish the difference between common versus uncommon varieties (or valuable minerals). The Act states:

> *. . . common varieties does not include deposits of such materials which are valuable because the deposit has some property giving it distinct and special value, and does not include so-called "block pumice" . . .*

Thus, it has been left up to the Department of Interior and the courts to decide what is common or uncommon under the law.

The Act also provides for the multiple use of the surface of mining claims. Prior to the 1955 Act, mining claimants had exclusive rights to the surface as well as the mineral estate within their claim.

Federal regulations

The U.S. Forest Service has issued regulations in the Code of Federal Regulations (36 CFR 228, subpart C). The regulations try to differentiate between common and uncommon mineral materials under the 1955 Act. In the FS regulations, salable, or common variety minerals are defined as:

> *. . . mineral materials which consist of petrified wood and common varieties of sand, gravel, stone, pumice, pumicite, cinders, clay, and other similar materials. Such mineral materials include deposits which, although they have economic value, are used for agriculture, animal husbandry, building, abrasion, construction, landscaping, and similar uses.*

The FS regulations rely heavily on the "use" of the mineral in question to help define it as either a locatable or salable mineral. As an example, a deposit of high-purity silica sand, which is used by a claimant in the production of a glass sand, would be locatable under the Mining Law. Conversely, this same deposit would be considered a salable mineral if the material were simply sold as a base-course gravel. The regulations go on to define industrial minerals which meet the definition of locatable minerals as follows:

> *. . . minerals used in manufacturing, industrial processing, or chemical operations for which no other mineral can be substituted due to unique properties giving the particular mineral a distinct and special value . . .*

The FS regulations list several types of minerals which generally fall into the locatable category, including, but not limited to:

minerals other than calcium or magnesium carbonate used to chemically alter the soil;

limestone suitable and used for cement manufacture, metallurgy, production of quicklime, etc.;

silica suitable and used for glass manufacture, silica flux, etc.;

alumino-silicate or clays having exceptional qualities suitable and used for the production of aluminum, ceramics, drilling muds, etc.;

gypsum suitable and used for wallboard, plaster, etc.;

block pumice which occurs in nature with pieces having one dimension of two inches or more and which is valuable and used for applications that require such dimensions, etc.;

stone recognized through marketing factors for its special and distinct properties of strength and durability . . . suitable and used for structural support.

The BLM regulations at 43 CFR 3601 also provide criteria to distinguish common variety from uncommon variety minerals. Salable, common variety minerals are defined in the BLM regulations as:

> *. . . deposits that may have a use in trade, manufacture and science, or in the mechanical or ornamental arts, but do not possess a distinct special economic value over the general run of such deposits.*

The regulations go on to characterize uncommon deposits in the following way:

If a deposit has distinct and special properties that make it commercially valuable in manufacturing, industrial or processing operations it may be uncommon. The commercial properties which may make a deposit uncommon can include: quantity of the deposit; geographical location; proximity to market or point of utilization; accessibility to transportation; reserves reasonable and consistent with expected demand; and feasible methods for mining and removal of the material.

Thus, the BLM regulations are more generic than the FS regulations and do not attempt to list or describe particular mineral types or mineral groups which fall into one category or the other. Also, the BLM regulations, unlike the FS regulations, include extrinsic factors of the mineral deposit such as geographic location and transportation accessibility when considering whether a mineral deposit is common or uncommon.

Each agency has an administrative process whereby industrial minerals are formally categorized as either locatable or salable. The FS prepares a "classification report" when trying to make this determination. This is an internal report used by FS decision makers. In a similar fashion, the BLM prepares a common variety determination as part of the standard, validity examination process. Each method is used to arrive at a decision point, namely, based on the specifics of the mineral deposit in question, should it be administered as a locatable or salable mineral under the various laws.

So, what's the big deal?

Table 1 displays some of the major advantages and disadvantages for industrial mineral producers with regard to whether or not a particular mineral deposit on Federal land is classified as a locatable or salable mineral.

Table 1. Advantages and disadvantages for classifying mineral deposits

Locatable	vs.	Salable
Approval of mining by the BLM/FS is non-discretionary		Approval of mining at the discretion of the BLM/FS
No royalty paid to U.S. government		Royalty paid to U.S. government based on fair market rates
Patenting right		No patenting rights
No sales by U.S. government		Either competitive or non-competitive sale
BLM doesn't require reclamation on areas less than 5 acres		Reclamation required on most operations by BLM
FS requires reclamation		FS requires reclamation

Interior Board of Land Appeals

The Interior Board of Land Appeals (IBLA) is the judicial branch of the Department of Interior. It conducts administrative hearings on appeals to decisions made by officers of the Department. Because the BLM has authority over mineral resources on Federal lands, appeals to decisions affecting the mineral estate on public lands and national forests are brought before the IBLA.

Many IBLA decisions have had an impact on the way the BLM and FS classify particular industrial minerals on Federal land. Two landmark cases heard before the IBLA, *U.S. Minerals Development Corp.*, 75 ID 127 (1968), and *McClarty v. Secretary of the Interior*, 408 F2nd 907, 908 (9th Cir 1969), set forth the standards to distinguish between uncommon and common variety minerals. The *McClarty* case, in par-

ticular, clearly established the criteria by which these deposits are to be judged. It involved the classification of an andesite building stone in Washington State that was found to be locatable based on its intrinsic properties which included natural fracturing and flat-surfaced cross sections. The following guidelines from the *McClarty* case are used by BLM and FS geologists and mining engineers when evaluating certain industrial minerals:

1) There must be a comparison of the mineral deposit in question with other deposits of such minerals generally;

2) The mineral deposit in question must have a unique property;

3) The unique property must give the deposit a distinct and special value;

4) If the special value is for uses to which ordinary varieties of the mineral are put, the deposit must have some distinct and special value for such use; and

5) The distinct and special value must be reflected by the higher price which the mineral commands in the market place.

Case studies

To illustrate some of the IBLA decisions and their impact on specific industrial minerals, several case studies are examined below.

Limestone

In many instances, limestone (and dolomite) is considered to be a common variety stone, particularly when you consider its wide geographic distribution within sedimentary rock terranes. In 1969, the Board held in *U.S. v. Chas. Pfizer & Co.*, 76 ID 331, that certain grades of limestone used in the manufacture of cement were locatable. The Board set a standard for limestone deposits, stating that they need to contain greater than 95% calcium or magnesium carbonate to meet the definition of an uncommon variety stone. Thus, limestone which is used in the manufacture of cement is locatable. On the other hand, suppose another limestone was used as a concrete aggregate. It is likely that the limestone in this case would be found to be common variety stone. This is due, in part, because other types of stone can be substituted for limestone as an acceptable concrete aggregate, whereas the chemical requirements for cement, metallurgical lime, and other specialty products are fairly strict and not all limestones can meet them. **Figure 2** is a limestone quarry located south of Tucson, Arizona, where part of the operation occurs on BLM land and part occurs on the Coronado National Forest. This mine, once operated by Pfizer, is now owned and operated by Specialty Minerals, a division of Mineral Technologies Incorporated. At the Specialty Minerals mine, both locatable and salable limestone are produced. The majority of the quarry rock is high-calcium limestone used for locatable purposes, including uses in the paint industry and also for desulfurization. The remaining portion of the deposit is mined for decorative rock products and sold in the local market area. The decorative rock is sold under a salable minerals contract and a fair-market royalty is charged to the company.



Figure 2. Mining of "locatable" high-calcium limestone and common variety stone at Specialty Minerals' operations south of Tucson, Arizona. The mine occurs on both BLM and FS lands.

Pumice

Deposits of pumice are relatively common throughout the western U.S. where Tertiary and Quaternary volcanic rocks are present. Pumice was one of the minerals specifically listed by Congress in the 1955 Surface Resource Act as a common variety mineral, with the exception of "block pumice," which occurs in nature in pieces having one dimension of two inches or more. However, in 1991, in *U.S. v. Multiple Use, Inc.*, 120 IBLA (1991), the IBLA held that an air-fall pumice used as an abrasive for stone-washing denim is an uncommon variety mineral. The Board went on to explain that three properties of the pumice deposit gave it a distinct and special value. Those properties are coarse particle size, particle shape, and absence of staining. The particular pumice deposit in that case was Arizona Tufflite's White Vulcan deposit, located outside of Flagstaff, Arizona. It was shown in that case that pumice, in pieces 3/4 inch (19mm) or more in diameter, is used to abrade denim in the stone-wash laundries in El Paso, Los Angeles, and elsewhere. Pumice meeting the requirements for stone-washing commands a premium price over and above the value of ordinary pumice.

Another pumice deposit, mined by Copar Pumice Company in New Mexico, has similar characteristics to the White Vulcan deposit (**Figure 3**). It has recently been classified by the FS as an uncommon variety mineral. At a typical pumice mine like this, particles of pumice in sizes 3/4 inch (19mm) or greater is screened and separated from pumice less than 3/4 inch in size and the two are treated as uncommon and common varieties, respectively, similar to the limestone example discussed above. The locatable portion of the deposit is removed under the Mining Law and the salable portion of the deposit is charged a royalty to the government.



Figure 3. Large-sized pumice fragments within the El Cajete Pumice of northern New Mexico, which is mined and used for stone washing in the garment-finishing business. Pumice sized greater than 3/4 inches (19 mm) in diameter is separated from other pumice and lithic fragments and sold at a premium price.

Decorative stone

Decorative stone products is another area of mineral case law that is still not very well understood. Two decorative stone operations in Arizona help to illustrate how the IBLA's *McClarty* criteria are typically used to classify materials as common or uncommon stone. In the first case, a decorative stone product called Sierra Stone is produced from several small quarries in east-central Arizona. The rock is a buff colored sandstone, with prominent Liesegang banding on cut surfaces. The iron-oxide banding creates "desert landscape patterns" along laminations and cross-bedding in the rock (**Figure 4**). The rock is often cored and cut into thin slabs for use as coasters and other novelty products. FS geologists were asked to classify the material in order to determine if it should be sold under contract from national forest land or if it is properly located as a mining claim. They determined that the unprocessed Sierra Stone sold at a much higher value than other sandstones and other types of rock used for common purposes. After conducting their analysis of the material, they concluded that it possesses a distinct and special value that is reflected in the price paid for the material in the market. Although sandstones are generally common in the southwestern U.S., only a handful of deposits exhibit the physical properties and pricing structure of Sierra Stone. This particular deposit is an uncommon variety and properly located under the 1872 Mining Law.

Apache Rainbow rock, which is also produced from small quarries in Arizona, is useful as a comparison to Sierra Stone. Apache Rainbow rock is a banded quartzite that when cut into slabs, exhibits very interesting patterns and variegated color schemes (**Figure 5**). The patterns and banding are due to fractures and sedimentary features

such as load casts and ripple marks that have been infilled with iron oxides and mafic minerals from intrusive rocks that have altered the host quartzite. This material was classified as a common variety mineral by a FS geologist. Although Apache Rainbow rock exhibits many of the decorative qualities that Sierra Stone does, its use in the marketplace is limited to landscaping and similar uses. Landscaping is one of the common variety uses listed in the FS regulations. More importantly though, a market analysis indicated that Apache Rainbow rock did not possess a distinct and special value that was reflected by a higher price in the market place. It was shown that Apache Rainbow rock sells in the same price range that other typical, common stone products such as granite, basalt, and marble, sell for. Thus Apache Rainbow rock did not meet one of the crucial tests in the *McClarty* criteria.



Figure 4. Coconino Sandstone with Liesengang banding. Cored rock is sawn into thin slabs for sale as coasters in novelty shops. "Sierra Stone," shown here, was classified as a locatable mineral by the FS mineral examiners.

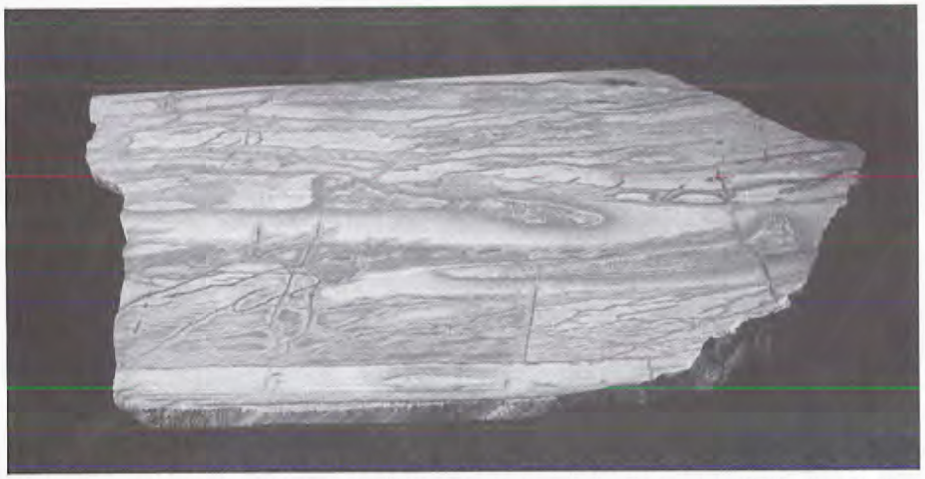


Figure 5. Dripping Springs Quartzite ("Apache Rainbow") displays variegated color patterns due to alteration with iron oxides and mafic minerals. FS mineral examiners determined this material to be a common variety stone.

Conclusion

Industrial minerals on Federal lands are classified as either locatable, leasable, or salable, depending on a variety of factors including physical and chemical properties of the mineral in question, uses of the mined products, and prices paid in the market place. Some industrial minerals easily fall into one of these three categories, while others are much harder to precisely define. In the latter case, each individual mineral deposit must be examined in light of the *McClarty* IBLA case criteria to help determine if it is a locatable or salable mineral.

Developing industrial mineral resources on Native American lands

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Abstract

Production of industrial minerals on Indian reservations is the fourth largest source of revenue on Native American lands. In 1982, the United States Congress enacted the "Indian Mineral Development Act" (IMDA). This law greatly expanded the participation of Native American tribes in the development of their energy and mineral resources. The evaluations of Indian Mineral Development agreements by the Bureau of Indian Affairs (BIA) are performed as assistance to the tribes entering these agreements and as a basis for determining if the agreements meet necessary criteria for Departmental approval. This paper will also discuss issues that are pertinent to doing business on Native American lands.

Introduction

Of the over 278 Indian reservations in the conterminous United States, only about 29 reservations do not have an identified sand and gravel resource. Reservations near major metropolitan areas are experiencing new interest in developing their sand and gravel resources. Several tribes have presented their mineral potential in various guidebooks and in the Bureau of Indian Affairs (BIA) - Division of Energy and Mineral Resources' (DEMR) publications. These tribes are interested in exploring the possibility of mineral agreements on their respective lands. The Division has also just completed a CD-ROM containing 111 regional geological reports compiled from publicly available data, covering over 278 reservations. In 1997 this CD-ROM will be available to all tribes as well as to major libraries and other government agencies.

The Congress enacted Public Law 97-382, the "Indian Mineral Development Act of 1982" (IMDA). This law greatly expanded the kinds of agreements that Native American tribes could enter into for development of their energy and mineral resources. It also revised the process by which these agreements are evaluated in fulfilling the Trust Responsibilities of the Secretary of the Interior. The Bureau of Indian Affairs provides assistance to the tribes entering into these agreements and determines whether the agreements meet necessary criteria for Departmental approval.

Background

Prior 1990 confusion existed over the status of sand and gravel as a mineral. Some tribes and BIA agencies treated sand and gravel as part of the surface estate and often developed sand and gravel under a business lease. Sand and gravel pits developed under such an arrangement generally did not have environmental assessments, mine plans, or reclamation plans and were poorly monitored, due to the confusion as to which regulations applied.

In 1990 the United States Claims Court issued an Order in the *Lyle Cochran, et al. v. United States*, Cl. Ct. No. 437-87L. The court held that Federal standards set forth in *James G. Watt, Secretary of the Interior, et al. v. Western Nuclear Inc.* Cl. Ct. No. 81-1686, (argued January 17, 1983–decided June 6, 1983) applied to this case and that sand and gravel was a mineral for the purposes of this case.

Based on these court decisions the Deputy to the Assistant Secretary-Indian Affairs (Trust and Economic Development) issued a memorandum to all BIA area directors defining sand and gravel as a mineral commodity. The 25 CFR Part 225 and draft CFR Part 216

(revised) also acknowledge sand and gravel as mineral commodities for the purpose of leasing and permitting. The regulations propose to set a 5000 cubic yard threshold above which will require a mining and reclamation plan.

Statutory authority

Legislation authorizing mineral development on Native American lands was first enacted on March, 1909, (35 Stat. 783; 25 U.S.C. 396) for individual Native American lands only, commonly referred to as allotted lands. The Act directed the Secretary of Interior to develop regulations for implementation of the legislation but provided minimal guidance as to form or content. The resultant regulations provided for leasing of lands for mineral development and evolved into more or less a standard form lease with variable provisions on bonus bid, acreage rental, royalty, bonding, etc.

The Act of May 11, 1938 (52 Stat. 347; 25 U.S.C. 396a et seq.) allowed mineral development on tribal lands. This Act provides that leases be limited to 2560 acres per lease and that duration of the lease would be for a primary term of ten years and so long thereafter as minerals are produced in paying quantities. Leasing regulations for allotted lands were subsequently amended to also include the acreage limitation and duration provisions. As a result, the lease forms for tribal and allotted lands are essentially the same.

The two leasing Acts generally limited participation of the Native American mineral owner in actual development and production activities. For the most part, bonuses, rentals, and royalties were the only opportunities for benefit from development of Native American energy and mineral resources. In addition, the ten-year primary term lease restriction, although workable initially, proved troublesome with the arrival of large modern-day developments. This was especially true for solid minerals, because the time required to make a discovery, complete exploration, and perform feasibility studies (including delays from market conditions), mine preparation, and construction of processing facilities often exceed a ten-year period. The lease could only remain valid beyond the ten-year primary term through continued production in paying quantities. Also, if an area of interest exceeded 2560 acres it required multiple leases that increased administrative burden to maintain the legal documents.

The Indian Mineral Development Act (IMDA), besides providing tribes opportunity to actively participate in mineral development, eliminated the shortcomings of the leasing Acts by making the forms of agreements, terms, and acreage negotiable. It should be noted, however, that the Act only authorized **tribes** the opportunity to enter into these agreements. Individual Native Americans can not make IMDA agreements. However, Native Americans holding trust property within the overall project area can include their mineral resources in a Tribal Minerals Agreement, provided the parties agree and the Secretary determines it is in the individual Native American owner's best interest.

The IMDA does not prescribe what form the agreement must take; how the parties will participate; whether there is a carried interest, a working interest, or both; what the term will be; how many acres will be involved; what employment and/or contracting opportunities may be negotiated; and how the products will be disposed of, i.e., sold, taken in kind, or some combination thereof. In other words, all terms and conditions of the agreement are negotiable. Consequently, each IMDA agreement is unique and all elements contained in the negotiated agreement must be evaluated by the Secretary of the Interior to determine if the agreement provides the best equity to the tribe for the contemplated development.

The Secretary must either approve or disapprove any complete agreement within 180 days of its submittal by the tribe. Information as to the character and extent of any possible ore body sufficient for a full environmental analysis is rarely available during the first stages of development. Since the National Environmental Policy Act (NEPA) must also be satisfied, initial approvals cover the economic terms only and are conditioned on complete compliance with NEPA requirements before land disturbance activities are initiated.

Regulatory authority

The following Codes of Federal Regulations (CFR) apply to leasing of solid minerals on tribal and allotted Native American trust lands:

25 CFR 211, Leasing of Tribal Lands for Mining (revised);

25 CFR 212, Leasing of Allotted Lands for Mining (revised);

25 CFR 213, Leasing of Restricted Lands of Members of Five Civilized Tribes, Oklahoma, for Mining;

25 CFR 214, Leasing of Osage Reservation Lands, Oklahoma, for Mining, Except Oil and Gas;

25 CFR 215, Lead and Zinc Mining Operations and Leases, Quapaw Agency;

25 CFR 216, Surface Exploration, Mining, and Reclamation of Lands;

25 CFR 217, Management of Tribal Assets of Ute Indian Tribe, Uintah and Ouray Reservation, Utah, by the Tribe and the Ute Distribution Corporation;

25 CFR 225, (Proposed) Oil and Gas, Geothermal, and Solid Mineral Agreements (draft); and

25 CFR 226, Leasing of Osage Reservation Lands for Oil and Gas Mining.

On November 21, 1991, the Department of Interior published draft rules for 25 CFR 211, 212, and 225 in the Federal Register. 25 CFR 211 and 212 were retitled and now read:

25 CFR 211, Leasing of Tribal Lands for Mineral Development; and

25 CFR 212, Leasing of Allotted Lands for Mineral Development.

BIA processing of IMDA agreements

The BIA - Division of Energy and Mineral Resources (DEMR) evaluates agreements as to the best interest determination related to potential economic return to the Native American mineral owner. Determinations related to environmental, social, and cultural effects on the tribe are made by the Bureau of Indian Affairs with personnel at area and agency offices. The Division reviews where lands under the proposed agreement are located, considers geologic and geographic features within the area, and makes comment on any concerns that may result from this review. These concerns may result from proximity to private residences or communities, other commercial developments, water resources, or alternative land uses.

If the approving official at the BIA agency or area office feels the proposed agreement(s) are deficient in the requirements necessary for approval and the tribe/developer disagree, the agreement(s) must be forwarded to the BIA central office in Washington, D.C. for review. If the central office sustains the findings of the agency or area office and the Assistant Secretary of the Interior for Indian Affairs disapproves the agreement(s) any further recourse must be decided in the District Court of the United States.

Commonly, tribes request the DEMR provide advice, counsel, and assistance either at the outset or at some stage in the negotiation of a Minerals Agreement. Generally this expedites the process because problem areas in a proposed agreement can be identified early in the process and corrected before the final document is submitted for approval by the tribe.

Mineral resource data on Native American lands

The Division of Energy and Mineral Resources maintains a comprehensive data base on Native American energy and mineral resources. In 1993, the Division developed the automated National Indian Energy and Mineral Resources (NIEMR) data base to house, maintain, and manage these data resources for the individual tribes.

The Division archives Native American resource information and can distribute published information to agencies and individuals who wish to retrieve data for review and further analysis. Unpublished Native American minerals economic data are proprietary to the individual owner and can be released only with written approval of the owner.

The Division has published a CD-ROM disk that can be read on desktop personal computers using a standard *Windows*® operating system. The initial CD-ROM for the NIEMR program contains published Mineral Assessment reports written by contractors, the U.S. Geological Survey, and the U.S. Bureau of Mines. These regional overviews of each reservation contain a geologic discussion of the reservation and a general assessment of the mineral resources, including base and precious metals, coal, and oil and gas. Each report contains a geologic discussion of the reservation, the structural and stratigraphic framework, and a general assessment of the mineral resources, including base and precious minerals, coal, oil and gas, and

industrial minerals, such as sand and gravel. Reports include correlation and stratigraphic charts, geologic cross sections, surface topographic and cultural maps, and subsurface structure and lithofacies maps. These reports were designed to compile and summarize publicly available information on the geology, mineral and energy resources, and the potential for economic development of all Native American reservation lands. The primary sources for this information were published reports, field studies, or other on-site investigations.

There are 111 reports on one CD-ROM. Included on the CD-ROM are complete report text, tables, figures, and maps. Also included with each report is an excellent bibliography, providing a great reference resource list for someone contemplating exploration or development on Native American lands. The organization of the books on disk is by reservation, but easy-to-use indices are provided for quick searching of all reports by either location (reservation, state, or area office) or by type of resource, such as oil or a specific type of mineral.

New mineral resource data and/or development opportunities

The Division of Energy and Mineral Resources has an "outreach program" which provides opportunities for tribes to present new data of the results of mineral assessment projects and the mineral resources discovered on their reservations to industry and the public. The Division of Energy and Mineral Resources works with tribes to assist in providing opportunities, in the form of marketing events, for tribes to promote development on energy and mineral prospects on their reservations. The Division utilizes national conferences and regional symposia to present to industry the results of Minerals Assessment projects and the mineral resources discovered on their reservations. The Division of Energy and Mineral Resources also sponsors an annual Indian Minerals Conference in Denver. Tribal representatives present poster displays at these conferences and meetings; and participate in the conferences, focusing attention on oil and gas, coal-bed methane, and other mineral resources on Native American trust lands.

Participation of Native Americans at these meetings stimulate interest on reservations by:

-Identifying Native American tribes that are interested in considering a mineral proposal.

-Discussing the positive effects of doing business on Native American lands.

-Establishing ground rules for mineral proposals and negotiations.

-Discussing successful exploration methods on Native American lands, thus aiding companies in the design of more efficient exploration programs.

-Publishing new geotechnical data on mineral resources on Native American trust lands.

Whenever a tribe requests assistance for promoting their energy and mineral resources, the Division works with the tribe to design a public meeting or conference. Technical papers describing the energy or mineral resources on the reservation are presented at a technical meeting attended by industry personnel. Tribal representatives present poster displays at these conferences and meetings; and participate in the conferences, attempting to focus attention on the resources on their reservation.

The Division advises any tribe interested in participating in the "outreach program" that while the purpose of the conference is to stimulate interest in developing Native American lands, participation in the outreach program by a tribe does not commit tribes in any way in leasing or developing any of their lands. The tribes, themselves, control all mineral leasing activities on their respective reservations.

Tribes that wish to have their mineral assessment information made public can have their data published. Published information is distributed to libraries, state geological surveys, and university geology departments. Thus their Native American energy and mineral resource information becomes available to any company conducting a literature search. Of course publication of Native American mineral information can only occur with the consent of the Native American owner.

Issues on Native American lands

Industry interested in doing business with Native Americans should be aware of sensitive issues on reservations. Protocol is also very important. It is recommended that anyone interested in doing business with Native Americans first contact the tribal chairman of the Indian reservation to establish a point of contact. This contact person can discuss the political organization of the tribe, how and when the Tribal Council

meets, tribal tax structure, tribal mining codes, and other factors that would impact mineral development. In addition, the contact person can describe the cultural considerations such as religions and traditions that may affect their decisions. Tribal culture varies from tribe to tribe, even within a small geographic area.

While the BIA has a trust responsibility to Native American tribes in the United States, on many Indian reservations the BIA's role is reduced and many functions are now being performed by the tribal government. Much of the BIA's and Division's current activities are focused on preparing the tribes for compacting their energy and mineral resource development activities. New programs have been initiated that will assist the tribes in taking total control of the management and development of their natural resources, and in improving their revenue return.

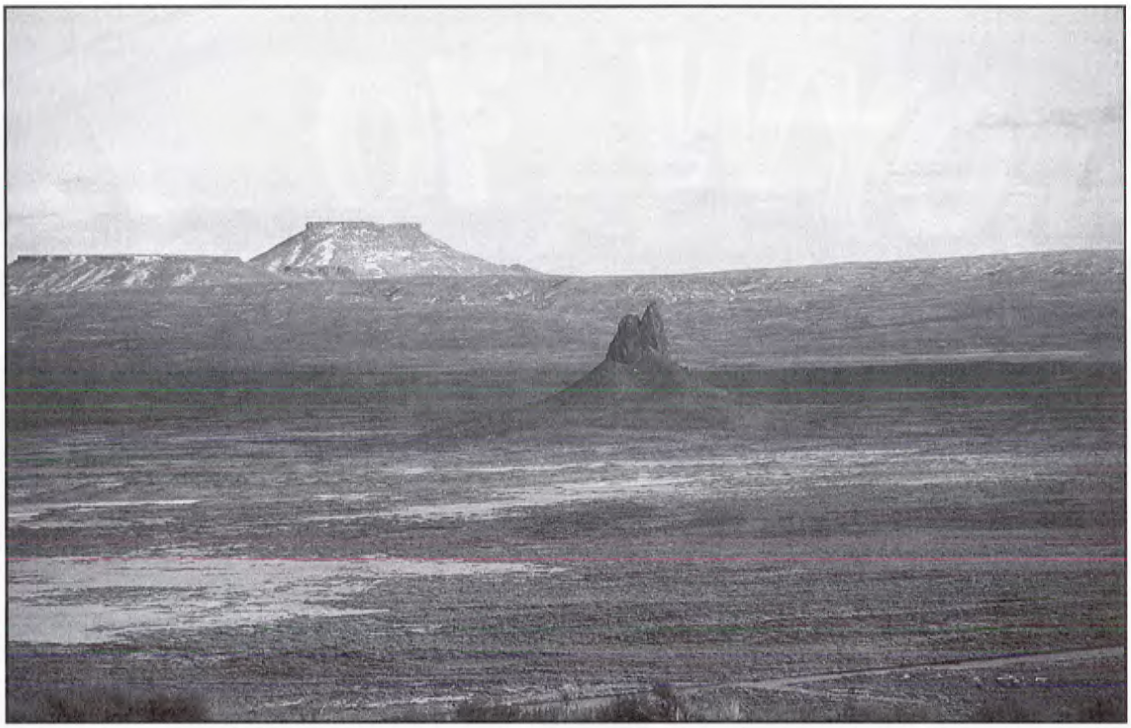
Some tribes have enacted additional tribal mining codes to supplement the Federal government regulations. These tribal regulations are intended to address specific needs of the tribe and are supported by the BIA.

Companies should be aware the sovereign immunity status of individual tribes. As a general rule, tribes cannot be sued without the consent of or waiver by the tribal government. Companies should research the area of tribal law and tribal courts in order to get an understanding of remedies available on a particular reservation. Tribes may even have to enact specific waivers for a particular situation in order to allow economic development on the reservation.

Conclusion

The Native American tribes retain broad sovereign authority to regulate activities within the external boundaries of their reservations. Companies that are successful in doing business with Native American tribes have been able to identify the unique cultural, governmental, and economic differences that apply on Indian reservations. In addition, the Indian Mineral Development Act offers an unique opportunity to tribes and industry by making the forms of agreements, terms, and acreage all negotiable.

The tribes benefit from attending professional meetings and exchanging ideas with energy and minerals personnel. Industry also gains new insights as it learns about the tribes' resources and concerns about mineral development. The political climate is changing radically with the introduction of Tribal Self Governance and the transfer of authorities from the Department of the Interior to the tribes themselves. However, for so long as the Trust Responsibility exists, the Department of the Interior, the Bureau of Indian Affairs, and the Division of Energy and Mineral Resources will look forward to opportunities to assist the tribes in promoting their own economic independence through development of their own mineral resources.



Boar's Tusk north of Rock Springs in southwestern Wyoming is a one million-year-old volcanic neck located on the edge of the Killpecker sand dune field. The white, vegetation-free areas are active parts of the dune field. The mesas in the background are primarily Eocene Green River Formation capped by one million-year-old volcanic rocks. This area is known as the Leucite Hills volcanic field. (Photograph by R.W. Jones.)

Living with rare and endangered species: A proactive approach for industry

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Abstract

Those of us in the mining industry who may be wishing that the present Republican Congress will do "magical things" with the Endangered Species Act (ESA) are going to be disappointed. The ESA is not going to go away. In fact, there is a 50-50 chance that it may be strengthened.

Can a mining company live with endangered species? The answer is yes, but only if a proactive approach is taken toward protecting the environment. A mining company today, regardless of size, must be alert and knowledgeable and willing to give and occasionally "turn the other cheek."

A mining company must be alert to the potential for existing endangered species on a property being considered for acquisition and also to the type and nature of our reclamation programs. These will often create habitat and possibly "introduce" endangered species to the site.

A company must be knowledgeable about the Federal ESA and the endangered species regulations in those states where their operations are located. A knowledgeable company will know the key players on the Federal and state levels along with any private or citizen environmental groups. Most important is that the company have, either on staff or employed as a consultant, a trained professional who is aware of the regulations pertaining to endangered species and is able to do an inventory on company sites. This professional would also act as communication liaison between key players in the Federal and state agencies and the company.

To "give" is to cooperate with those agencies charged with endangered species protection. Cooperation can occur in a variety of ways. To "turn the other cheek" is to be able to lose a decision regarding an endangered species but not lose sight of environmental awareness. This must continue if the company hopes to succeed in future years.

Introduction

Today, business and nature must be able to mix if both are going to succeed. The extractive resource industries are missing a great opportunity by not taking a proactive approach to the environment. This in turn will translate into better relations with their communities, neighboring residents, state and Federal agencies and, yes, environmental activists. A good place to start a proactive stance is in wildlife enhancement by using some innovative reclamation methods. For the purpose of this discussion, "wildlife" refers not only to fish and game, but also to plant habitat and species. Fish and wildlife species are regulated under the Federal Endangered Species Act (ESA). Plant species are also regulated under the ESA, but not with the efficacy that fish and game species enjoy.

The Endangered Species Act: The crown jewel or the pit bull of environmental laws?

In 1973, Congress enacted the Endangered Species Act (ESA). Now, twenty some years later, ESA is in debate in Congress over its reauthorization. Four "new" bills have been introduced which range from

property rights protection, compensation for "takings," and requiring government to consider the social and economic consequences of the Act, to a more moderate revision. This moderate approach calls for the extensive use of science in the listing of species and reduction of some Federal control. All four bills were introduced by Republicans in September, 1995. The most likely outcome for ESA reform in the current Congress will not be too "watered down" from the original ESA. What is taken out of the original Act may very well be replaced by state regulations.

The ESA still is an important Federal environmental statute and this also extends into many individual state endangered species regulations.

The most important aspects of the ESA are the provisions for land use. These rules are the "notorious" Section 7 (consultation) and Section 9 (taking).

Although Section 7 was originally intended as a restriction only on Federal agencies, its impact found the way into any project requiring Federal approval or involvement whether on public or private lands. This section requires each Federal agency, in consultation with the Secretary of the Interior, to ensure that any action it approves will not jeopardize the existence of any endangered or threatened species or **result in the destruction or adverse modification of habitat**. This is what surface mining companies must do to extract their targeted resource.

The "takings" Section (9) of the ESA was primarily intended as a prohibition against direct harm; i.e., shooting, but over the past twenty years, the courts have interpreted it to forbid private as well as Federal actions that destroy habitat. It is this "take" provision which has the greatest potential to limit private property use and has made the ESA a primary obstacle to land development in this country.

What is a "take"? According to ESA, "'take' means to harass, harm, pursue, hunt, shoot, wound, kill, trap, capture, or collect ..." Because the ESA did not define "harass" or "harm," the Fish and Wildlife Service (FWS) did:

> *"Harass" in the definition of "take" in the Act means any intentional or negligent act which may create the likelihood of injury to wildlife through annoyance which would then disrupt their normal behavioral patterns which can include, but is not limited to, breeding, feeding, or sheltering.*
>
> *"Harm" in the definition of "take" in the Act means an act which actually kills or injures wildlife. This can mean the significant habitat modification or degradation where it can lead to the actual killing or injuring of wildlife.*

The FWS regulatory definition of "harm" is the one which creates the greatest concern for mining companies and other landowners. It explicitly includes the effects of habitat modification or land use actions, such as clearing trees or brush to prepare a mining or plant site or, in some cases, reclamation activities which are usually required by local or state permit conditions. Recently, the courts have decided that "harm" can be established by proof of habitat modification alone, **assuming** "actual injury." Mining companies with land ownership in areas occupied by listed species are vulnerable to whether their land use actions will be viewed as "takings." This is where a proactive approach by the landowner becomes essential.

A mining company must be alert to the potential for the presence of endangered species, both plant and animal, on their property. Properties being considered for acquisition should be subject to a "wildlife" audit before purchase. Reclamation programs should be evaluated as to the type of habitat they might create. Some reclamation plans could create habitat that may induce endangered species to the site.

Usually, those endangered plant species occurring on private lands are regulated primarily by state agencies, while those that occur on Federal lands; i.e., mining claims, leases, etc., are regulated by the ESA or Federal agencies (e.g., U.S. Bureau of Land Management, U.S. Forest Service).

An example of the complexity of having to deal with listed species occurred on the Unimin mining operation at Byron, California. We knew that we had the Federal and state listed kit fox habitat on our properties and that we were required to mitigate for habitat destruction (taking) due to our mining.

At one point, we thought we had an agreement with the U.S. Fish and Game (USFG) that our mining/reclamation activity would be considered only "**temporary** habitat destruction;" however, our primary lead person with the USFG was transferred to Washington and the U.S. Fish and Wildlife Service (FWS) staff members who inherited our project disagreed with the position previously taken. They determined that our mining/reclamation activity was "**permanent** habitat destruction." As a result, we saw our mitigation for habitat compensation jump from a ratio of less than 1:1 (one acre habitat for every one acre disturbed) to a ratio of 3:1.

After about one year of meetings and negotiations with both the U.S. and the California Fish and Game

agencies, an agreement was reached which essentially resembled the original agreement. A Memorandum of Agreement was signed with the California Department of Fish and Game as the lead agency. FWS was able to concur with the California authorization. The fact that the Attorney General of California gave an opinion that "taking" of habitat under the California ESA is not a "taking" as defined by FWS played a major role in reaching a mitigation agreement with both agencies.

"Mitigation" is becoming the accepted form of accommodative flexibility under the ESA. A company can learn to live with the Act by voluntarily offering mitigating measures to avoid or offset any projected disturbances to habitat. By taking a proactive stance with the regulating agencies, a middle ground between "either/or" can be found. Rhetorical bad-mouthing of a species or agency is a waste of time and often counterproductive. Unimin was successful in negotiating an agreement with the respective agencies in California because we were knowledgeable about the species we had on our property, the Federal ESA, and perhaps most important, the "workings" of the California regulations pertaining to endangered species. We also made an effort to become acquainted with all of the key players on both the Federal and state levels. We also made the first move in setting up meetings and plant site visitations.

Getting along with regulatory agencies

Probably the first thing to understand about regulatory agencies is that they are made up of individuals of varying disciplines, temperaments, and philosophies. If a company approaches an agency well prepared, flexible, and ready to achieve some rational problem-solving that all parties can accept, dealing with the agency will not be distasteful or adversarial.

To achieve the above, a company must employ a proactive approach. Usually a company will have to interact with a regulatory agency if (1) the company wants something from the agency, or (2) the agency wants something from the company.

A good start to a proactive approach is to know all the key players in the agencies which a company is likely to interact with, along with any private or citizen environmental groups in the immediate area of the facility. The way to do this is to have a trained professional on the environmental staff who is aware of the regulations pertaining to endangered species and who has the educational background to do site inventories. This individual can be an in-house staff person or an outside consultant. Many companies have both, but agencies tend to be more impressed when an environmental person is part of the in-house staff. A trained in-house specialist will also function as a communication liaison between the key players in the Federal and state agencies and the company and as a "contact" person with private or citizen groups.

Once communications have been established, it is important to keep these lines of communication with the appropriate personnel open on an ongoing basis. Don't wait for a problem to occur. Dealing in good faith and solving issues in a spirit of reasonable compromise will establish a good track record. The regulatory agency and private groups will see the company as a responsible corporate citizen, but not a pushover.

Prior to launching into a permit situation, an intensive research effort is made to identify all the issues, strategies, and negotiating positions. The regulatory agency is not approached until all the research is completed. A good operating procedure is to ask the agency for an informal initial meeting. At this meeting, a company must be aware of how important it is to listen and be flexible to demonstrate to the regulatory agency that the company is interested in respecting the regulations, restrictions, etc.

This proactive approach to instigating the first meetings with an agency will set the standard for long term relations and will often result in saving money, avoiding delays, and minimizing the legal headaches. Also, it is good basic public relations and better for business because employees and local constituents are more enthusiastic about a pro-environmental company.

Many companies, even today, try to take a "defensive" stand against a regulatory agency. For example, they may plead for sympathy because they are only "a small company that really does not understand about environmental regulations." The agency's typical response may be that a heavy fine will prompt the company to get better educated about environmental compliance. Or, a company may make the argument for lenience because "we only did it once." This will have the same effect as the "babe-in-the-woods" defense. In most cases, a mining company will end up spending about the same amount of money whether they try to fight the restrictions or comply with them.

Another way to impress an agency is through cooperation by volunteering mitigation or supporting (usually financially) inventory studies or basic research projects either on or near the facility. Because most

Federal and state agencies are poorly funded for such work, this has a favorable impact on everyone.

An example of a cooperative research effort was a project undertaken by Unimin at its Emmett, Idaho facility. This research actually kept a nominated Federal Category 1 plant species from being listed as an endangered species. The plant, Aase's onion, occurred on mining claims located by Unimin.

For a period of three years, Unimin signed cooperative agreements with the U.S. Bureau of Land Management to research life history and habitat ecology and conduct regional surveys of the Aase's onion on both Federal and private lands.

Although Aase's onion is still a Federal Category 1 Candidate for Listing, Unimin's efforts revealed that onion populations were larger and more widespread than had been previously thought. It was also demonstrated that the species can be transplanted in reclaimed areas. Because of this research, it will eventually be removed from the Federal candidate listing.

Getting along with communities and neighbors

Mining companies must begin to enhance their public image, especially if they have operations in rural settings. Public opposition can be one of the most costly events leading to obtaining permits. When local communities become irritated or frustrated with a company, they will generally turn to national environmental and public interest organizations for guidance and help in opposing a company's plans.

Companies can take a proactive approach to begin setting up lines of communication with their community and neighbors. One way is to provide access to the community in order to remove any secrecy about the operations. Ordinary people are interested in mining and curious to find out what happens to the raw materials. Conducting an open house on Earth Day is an excellent beginning. Another is to set up a "neighborhood" committee which regularly meets with representatives of the facility to discuss any plant site or mine expansions, future permit requirements, and even to view future and current reclamation plans. All of the above will defuse any anxiety among the residents about an operation.

Another good public relations incentive is to be sensitive to residents concerns or complaints in a timely manner. Follow complaints up with an immediate phone call and, if necessary, a one-on-one visitation. It has been Unimin's experience that this approach has converted potential enemies into community partners with the company.

Probably one of the best methods of community involvement is through habitat enhancement. Mining companies still have the reputation for destroying habitat and leaving "ugly" scars. Unimin has taken a proactive approach in habitat enhancement. It begins by doing reclamation in a creative and ecological way that provides positive visual qualities and meets the permit conditions. Often it requires going above and beyond permit conditions. Plant site beautification is also essential because this area is what the public usually sees first. Most mine operations are hidden from the general public.

Habitat enhancement consists of any project, large or small, that protects wildlife. Its cost will vary with the intensity of the project, from merely allowing land to revert to its natural state, to planting trees, ground cover, and constructing lakes and ponds around a plant or mine site. Habitat enhancement creates environmental awareness not only in the community, but also among the company's employees. Many of these projects can be employee motivated with volunteer help.

There is a non-profit organization whose mission is to help companies inventory both plant and animal wildlife already present on corporate lands, and then help the company design a plan to enhance the habitat to provide shelter for those species. This organization, Wildlife Habitat Council (WHC), which was founded in 1987, has helped mining companies convince communities that mining operations can be compatible with wildlife. With the help of WHC, companies learn to implement effective wildlife management programs and, thus, convey a positive message to the surrounding communities. More important, WHC corporate members benefit from the publicity generated by the organization and from publicity on the local level through newspaper articles.

Conclusions

Representatives from the U.S. Fish and Wildlife attending a habitat enhancement meeting in Washington, D.C. in 1995, made it clear that a pro-environmental approach by industry is going to be great for business in the 1990s. The mining industry has an opportunity to create new ideas that are pro-business and environ-

mentally sound. As the debate over the future of America's wildlife continues, companies must adopt proactive alternatives to the reactive forces of Federal and state legislation.

Our industry must be willing to accept the fact that the American public favors strong environmental protection and is against any attempts to weaken the nation's environmental laws.

Finally, mining companies must be prepared to "turn the other cheek" if things don't go well for them on the environmental front. Losing a decision regarding an endangered species is not an excuse for losing sight of the positive effects of environmental awareness and of being an environmentally sensitive company. If the mining industry wants to succeed in future years, it must adopt a proactive environmental awareness stance.

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Presentations as abstracts only

Geology and economic development of the Piceance Creek Basin nahcolite

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The Piceance Creek Basin, located in northwestern Colorado, contains the world's largest known deposit of nahcolite, a rare naturally occurring sodium bicarbonate mineral (commonly referred to as baking soda). In the north-central Piceance Creek Basin area, the Parachute Creek Member of the Eocene Green River Formation hosts bedded and disseminated nahcolite, associated with oil shale. In 1991, White River Nahcolite Minerals (WRN) became the first company to commercially mine nahcolite by combining horizontal directional drilling and hot water solution mining technology. The horizon currently mined is a 24-foot-thick section of the Boies bed, approximately 2000 feet below the surface. Hot water dissolves the nahcolite as the water passes along a horizontal hole drilled near the base of the Boies bed. The hot water is returned to the surface and subsequently cooled, casing dissolved sodium bicarbonate to recrystallize. WRN produces high purity sodium bicarbonate, which is sold into the estimated 500,000 ton-per-year U.S. pure, food grade, industrial, and animal feed markets.

Precambrian geology of the Wyoming province

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The Precambrian geology of Wyoming is anchored by the Wyoming province which is an Archean craton composed of rocks that represent crustal-forming and crustal-accretion processes ranging from very early differentiates to the formation of greenstone belts. In response to high rates of heat flow, and smaller, more numerous and more active mantle convection cells which dominated Archean time, the province contains greenstone belts, mobile belts, and high-grade supracrustal sequences which are interposed in granite-gneissic terranes. Within the state, Proterozoic rocks, seen at the edges of the Archean craton, are characterized by the occurrence of major, Early-Middle Proterozoic mobile belts and accreted island-arc systems which abut and are welded to the Wyoming province. These accreted terranes are intruded by 1.8 to 1.6 Ga (billion years) granitic and mafic complexes, and during Late Proterozoic time, by "anorogenic-type" plutonic rocks comprising granite-anorthosite chemistries.

The Archean rocks which form the Wyoming province are dominantly Middle to Late Archean gneisses and migmatites. Although no Early Archean rocks have yet been identified, zircons, separated from supracrustal masses within Middle Archean gneissic terranes (Beartooth Mountains), yield crystallization dates of about 3.9 Ga and as such, record the existence of some of the world's oldest known crustal sequences. Gneissic rocks (Wind River Range and Owl Creek Mountains) of Middle Archean age are of uppermost amphibolite grade and locally reach granulite-grade metamorphism. Temperatures occurring with these grades have generally reset most isotopic clocks to the 3.6 to 3.0 Ga range. Potassium-poor gneisses (Sierra Madre, Granite and northern Laramie Mountains) yield ages in the 3.0 to 2.8 Ga ranges and relict ages of crystallization of ~3.2 Ga. As a group they reflect a regional thermal event which probably reset the ages of late Middle Archean sequences. Potassium-rich intrusive systems are the dominant

rock types of the Late Archean (central Laramie Mountains).

Throughout Archean time, rifting within the craton–and perhaps along its edges–sponsored the formation of greenstone belts, iron formations, and very early fining-upward sedimentary sequences. Such troughs were often short-lived and resulted in the preservation of only pieces of the "standard" section for greenstone belt-style sediments and supracrustals. Such failed rifts, upon closure, very quickly metamorphosed their components to relatively high grades. As such they are called "high-grade supracrustals" (Tin Cup District, Granite Mountains, Owl Creek Mountains). Complete greenstone belt sequences developed in Late Archean time (2.8 to 2.7 Ga) (South Pass, Seminoe Mountains, central Laramie Mountains) with basal ultramafic rocks, pillow basalts, and sedimentary rocks. At the end of Archean time (2.7 to 2.5 Ga) granites intruded and folded greenstone belts and may have remobilized gneisses (South Pass and central Laramie Mountains).

Archean rocks of the Hartville uplift in the eastern part of the state do not fit the tectonic style or petrologic assemblages of the Wyoming province. Rocks here record the interlayering of basaltic flows and pillow lavas with stromatolitic photic-zone carbonates, which suggests that this part of the craton hosted a relatively quiet site of marine deposition with volcanic activity in close proximity. Archean granitic gneisses (2.6 Ga) of the area are locally intruded by and/or tectonically interleaved with granites yielding Middle Proterozoic ages (2.0 to 1.7 Ga). These field relationships and geochemical data showing the occurrence of S-type granites, anatexis, and isotopic re-equilibration suggest an alternate possibility. Because the Trans-Hudson Orogen (1.8 to 1.6 Ga) of central Canada (here considered to be a complex major mobile belt) extends through the eastern edge of Wyoming, the poorly exposed rocks of the Hartville uplift may therefore represent a relatively undeformed portion of an exotic terrane transported into its current position during that event. The dynamics of transport would easily generate melts and interleave slivers of different terranes.

Proterozoic terranes of Wyoming are best exposed at the southern edge of Wyoming (southern Laramie Mountains, Medicine Bow Mountains, and Sierra Madre). Lying unconformably on the edge of the Archean cratonic block is an Early Proterozoic-type mature clastic wedge (2.7 to 1.8 Ga). This section, containing the Snowy Pass Group and the Phantom Lake Metamorphic Suite, is a transgressive sequence of fluvial and shallow marine, mature clastic to deep-water black shale sediments (Medicine Bow Mountains, Sierra Madre). Thrust obliquely against this section, and thus forming part of a crustal suture called the Cheyenne belt, is the slightly younger (~1.8 Ga) Colorado province. The Colorado province is a sequence of volcanogenic rocks with island-arc affinities that represent a series of island-arcs which collided against the Wyoming province. The arc system volcanics of the Colorado province are intruded by slightly younger layered mafic and ultramafic complexes and mafic plutonic and granitic rocks (Lake Owen Complex, Elkhorn Gabbro, and Sierra Madre Granite) with ages ranging from approximately 1.8 to 1.6 Ga.

Intrusions which have been called "anorogenic" characterize the events of the Late Middle Proterozoic time. The Laramie Anorthosite Complex and the Sherman Granite (Laramie Mountains) date from about 1.5 to 1.4 Ga and represent the last major events in the Precambrian section of Wyoming.

Diamondiferous breccia pipes in the southern Green River Basin, Wyoming

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Several mafic to ultramafic breccia pipes and dikes were recently discovered in the vicinity of Cedar Mountain in the southern Green River Basin of southwestern Wyoming. The pipes lie along the margin of a

1000 square-mile anomalous region containing mantle-derived 'kimberlitic' indicator minerals. These detrital minerals have been recovered from ant mounds in Wyoming, the Bishop Conglomerate (Oligocene) in Wyoming and Colorado, and stream sediment samples along the northern flank of the Uinta Mountains of northern Utah.

The Green River Basin forms a topographic and structural basin containing surface exposures of Cretaceous and Tertiary sedimentary rocks. Based on oil and gas drilling, the basin near Cedar Mountain contains approximately 22,000 feet of Phanerzoic sedimentary rocks which overlie the Archean craton. The breccia pipes were emplaced along a north-south, 5- to 10-mile-long lineament expressed in the Bridger Formation (Eocene).

The cratonic rocks are completely buried and are not exposed in the basin except as xenoliths and associated xenocrysts in the breccia pipes. Other mantle xenoliths and xenocrysts occur in 3 to 1 Ma (mega-annum or million years) lamproitic breccias and flows in the Leucite Hills 60 miles northeast of Cedar Mountain. Mantle material, including diamonds, are also found in several Early Devonian kimberlite pipes, blows, and dikes in the State Line district of the Laramie Mountains, 200 miles east of the Cedar Mountain pipes.

Richard Kucera's discovery of 10 breccia pipes and associated dikes hosting mantle indicator minerals explains the source of many of the detrital indicator minerals in southwestern Wyoming. However, the pipes can not account for all of the indicator minerals in this region and the possibility of other undiscovered pipes in the Green River Basin and the Uinta Mountains is considered.

Dikes have been traced for thousands of feet and are up to 3 feet wide and locally enlarge into 30- to 50-foot-wide blows. The largest diatreme in the Cedar Mountain group measures 400 by 700 feet. The breccias are greenish gray and contain abundant lithic fragments from sedimentary units underlying the area, as well as crystalline basement rocks and eclogite nodules derived from the mantle. In addition, the breccia contains chromian diopside and pyrope xenocrysts.

Because of the abundance of crustal xenoliths, good xenolith-free samples are difficult to obtain for analysis. Analysis (on a weight-percent basis) of samples with apparently minor xenolith content range from 44.08 to 67.66% SiO_2, 0.27 to 0.34% TiO_2, 5.25 to 6.77% Al_2O_3, 2.82 to 5.13% Fe_2O_3, 0.05 to 0.22% MnO, 4.29 to 14.95% MgO, 5.10 to 17.99% CaO, 0.57 to 1.75% Na_2O, 1.41 to 2.20% K_2O, 0.26 to 0.56% P_2O_5, 6.6 to 17.77% LOI (loss on ignition), 0.07 to 0.14% Cr_2O_3, and 319 to 755 ppm (parts per million) Ni. The MgO content is low and the SiO_2 content is high compared to the average kimberlite, and the chemistry does not appear to be comparable to lamproite. Studies on the chemistry and petrography are continuing.

Guardian Enterprises, Ltd. recently completed a six-hole drilling program on the DK pipe. A total of 1250 feet of core was recovered. An additional 3000-pound bulk sample was collected near the surface of the DK pipe and from the dike complex. This bulk sample has been processed by the Wyoming State Geological Survey to recover indicator minerals for microprobe studies at the University of Arizona, and will be reprocessed in an attempt to recover diamonds on the Survey's grease table.

According to Guardian Enterprises, Ltd., three diamonds have been recovered from the property to date. One stone, with a maximum diameter of 2.2 mm, was found on the ground adjacent to one of the pipes. The diamond had SI clarity and J color. Two other diamonds were recovered from a 200-pound core sample by the Saskatchewan Research Council. The diamonds were reported to have maximum diameter of 1.07 mm and 0.13 mm, respectively. Other diamonds have been reported in the area in past years.

Estimating industrial mineral reserves in the real world

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The definition of reserves for industrial mineral (IM) deposits must formally meet the same requirements as for metallic or fuel minerals; the Securities and Exchange Commission (SEC) defines reserves as:

That portion of a mineable deposit which could be economically and legally extracted at the time of the reserve determination.

Yet IM reserves are commonly estimated with considerably less drilling and testing and at much less expense. This is mostly true because: (1) industry standards are commonly less prescribed for IM commodities, and (2) most IM operations cannot bear the heavy load of capital costs that intense reserve drilling would bring in the development phase of a new operation.

IM reserve estimates are generally based on wide-center drilling, followed by closer-centered drilling within a few years of anticipated production. Such secondary drilling provides mining control as well as more closely defined reserves, and may be written off as an operating expense. This approach may also be practical where product-lines or changing technology tend to change how the reserves are defined.

IM reserves defined with very few or no drill holes may be considered valid where geological continuity and grade/quality parameters can be prudently established. However, most often this approach is a clear formula for disaster.

Drilling and surface sample testing may not be enough to truly "prove" IM reserves. Commonly, the confirmatory steps of: (1) raw material characterization, (2) bulk sampling, (3) pilot plant testing, and (4) end-user testing, should follow drilling and surface sample testing to prove processability and marketability. Finally, mine planning and feasibility studies should be conducted to establish the profitability of extracting the material, whether the effort is to extend a reserve position or to prove reserves for a new venture.

World soda ash demand outlook relative to the Wyoming trona industry

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Next year will commemorate the 50th anniversary of the first shaft sinking by the Westvaco Chemical Corporation in the trona deposit of the Green River Basin. This was the beginning of the Wyoming soda ash industry. During the last half of the twentieth century, four more companies recognized the potential of this valuable resource and constructed mines and refineries in the region. Collectively, these five companies now represent about one third of total world soda ash production.

Stagnant market conditions that affected domestic soda ash consumption throughout most of the 1980s resulted in reduced revenues despite the increase in U.S. soda ash exports, which rose 167% during this time period. This situation, and the reevaluation and restructuring of core business components by the parent corporations, caused all five companies to ultimately sell all, or part of, their operations through joint venture partnerships to foreign firms that primarily produce or consume soda ash. These partnerships contributed to the significant increases in U.S. soda ash export sales during the past 15 years and the closures of a few of the partners' smaller, older synthetic soda ash facilities, thereby reducing world production capacity and soda ash availability.

Growth rates for soda ash consumption often are lower in most developed nations because of the cyclic nature of this mature commodity. The length of these cycles may vary, and is influenced by several factors, including price and availability of competitive caustic soda, use of glass cullet, reformulation of products that normally contain soda ash, and general economic conditions. These factors often make it difficult to accurately forecast long-term demand. In lesser developed nations where demand may be strong, the availability of hard currency, unstable social and political situations, high import duties to protect domestic soda ash producers, and lower per capita spending for consumer goods, may affect the quantity of soda ash imports. Despite these obstacles, several major international glass, chemical, and detergent companies with many worldwide manufacturing locations will construct new facilities in lesser developed countries that will rely on imported high-quality raw materials such as Wyoming soda ash. Some nations with large populations, such as China and India, continue to construct new soda ash plants or increase capacity at existing ones. These areas represent tremendous opportunities for increased soda ash exports; however, restrictive trade practices and the initiative to become self-sufficient in soda ash production capabilities impedes this effort.

Foreign participation in the Wyoming trona industry has provided the capital necessary for some of the

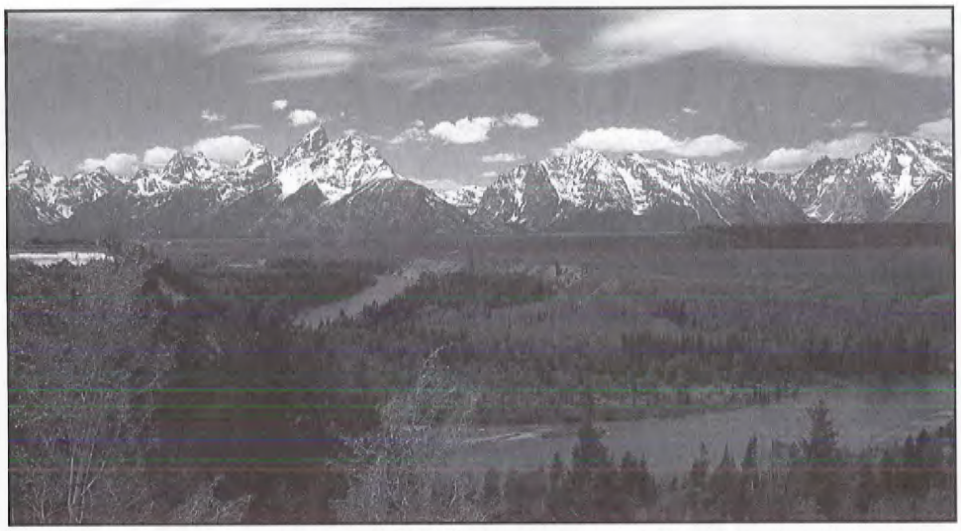
recent expansions. These expansions are important to maintain a long-term and reliable source of inexpensive natural soda ash. Changes in ownership, new joint ventures, plant expansions, and possible new grassroots facilities, will probably continue for the remainder of this century. Some of the lesser developed countries of today may become future joint venture partners in the twenty-first century as they become more prosperous and economically stable.

Wyoming trona — an overview of the geology

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Trona, an evaporite mineral of the compound sodium sesquicarbonate ($Na_2CO_3 \cdot NaHCO_3 \cdot 2H_2O$), was precipitated in a restricted lacustrine environment during deposition of the Eocene Wilkins Peak Member of the Green River Formation in southwestern Wyoming. Textural variations and sedimentary depositional features of the 25 bedded trona deposits are suggestive of multiple depositional scenarios.

Composed of 70% sodium carbonate, trona is referred to as natural soda ash, and refined soda ash is the major industrial product. Over 90% of the U.S. soda ash is refined by the five Wyoming trona producers. The total resources of the trona deposits is estimated at 127 billion tons. Of this, 40 billion tons of economical trona reserves are mineable by mechanical mining methods: conventional "hard rock" methods, mechanized extraction (drum-miners, boring machines), and hydraulically supported longwall shearing. Future use of secondary recovery through solution mining methods will substantially increase these reserves.



The Teton Range in Grand Teton National Park, northwestern Wyoming, rises abruptly from the valley floor of Jackson Hole. View to west across the Snake River. Grand Teton at 13,772 feet above sea level is the second highest mountain peak in Wyoming. Mount Moran is the high peak on right side of photograph. (Photograph by R.W. Jones.)

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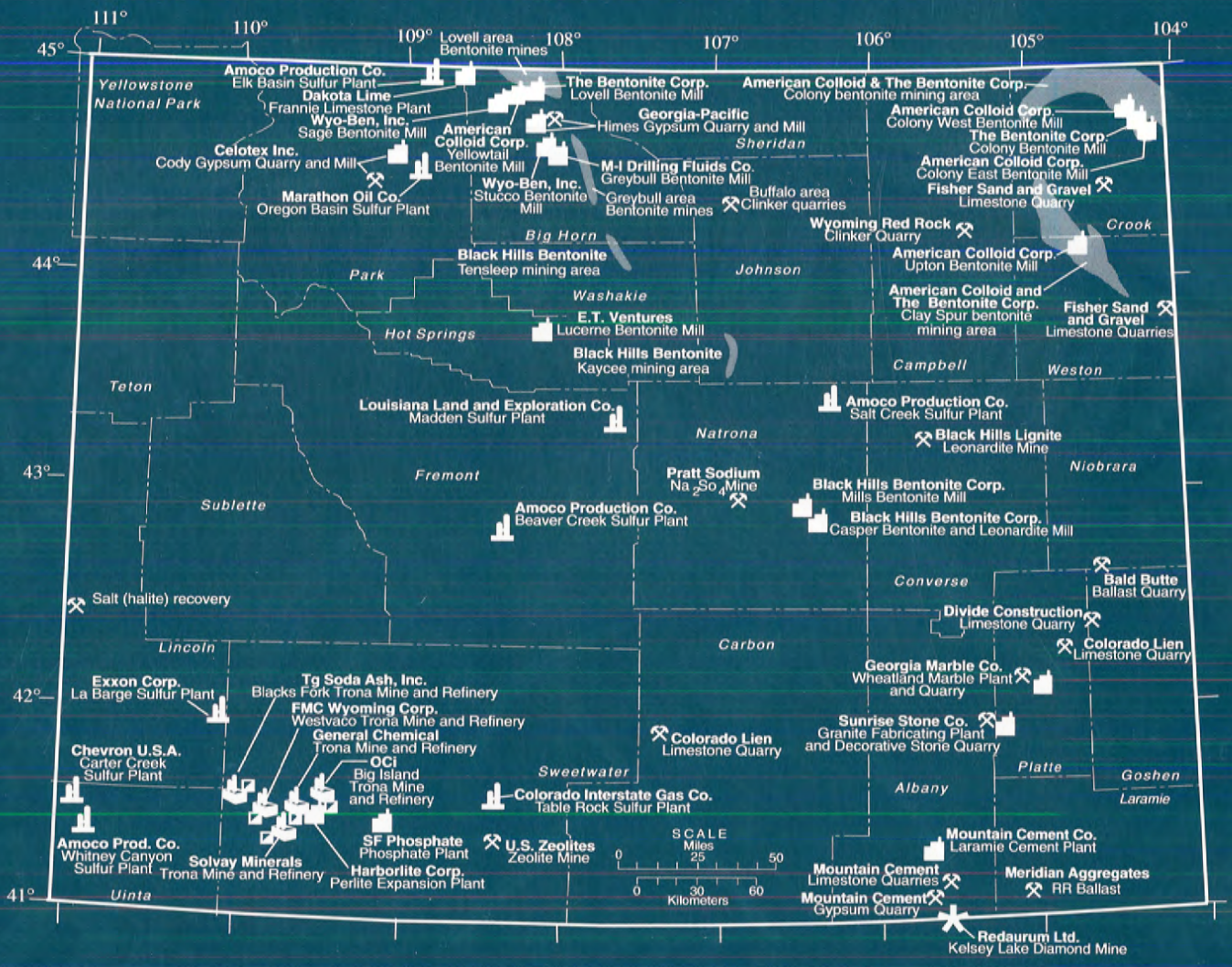
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Selected conversion factors*

TO CONVERT	MULTIPLY BY	TO OBTAIN	TO CONVERT	MULTIPLY BY	TO OBTAIN
Length			**Pressure stress**		
inches, in	2540	centimeters, cm	lb in^{-2} (=lb/in^2), psi	7.03×10^{-2}	kg cm^{-2} (=kg/cm^2)
feet, ft	3.048×10^{-1}	meters, m	lb in^{-2}	6.804×10^{-2}	atmospheres, atm
yards, yds	9.144×10^{-1}	m	lb in^{-2}	6.895×10^{3}	newtons (N)/m^2, N m^{-2}
statute miles, mi	1.609	kilometers, km	atm	1.0333	kg cm^{-2}
fathoms	1.829	m	atm	7.6×10^{2}	mm of Hg (at 0°C)
angstroms, Å	1.0×10^{-8}	cm	inches of Hg (at 0°C)	3.453×10^{-2}	kg cm^{-2}
Å	1.0×10^{-4}	micrometers, µm	bars, b	1.020	kg cm^{-2}
Area			b	1.0×10^{6}	dynes cm^{-2}
in^2	6.452	cm^2	b	9.869×10^{-1}	atm
ft^2	9.29×10^{-2}	m^2	b	1.0×10^{-1}	megapascals, MPa
yds^2	8.361×10^{-1}	m^2	**Density**		
mi^2	2.590	km^2	lb in^{-3} (=lb/in^3)	2.768×10^{1}	gr cm^{-3}(=gr/cm^3)
acres	4.047×10^{3}	m^2	**Viscosity**		
acres	4.047×10^{-1}	hectares, ha	poises	1.0	gr $cm^{-1}sec^{-1}$ *or* dynes cm^{-2}
Volume (wet and dry)			**Discharge**		
in^3	1.639×10^{1}	cm^3	U.S. gal min^{-1}, gpm	6.308×10^{-2}	1 sec^{-1}
ft^3	2.832×10^{-2}	m^3	gpm	6.308×10^{-5}	m^3 sec^{-1}
yds^3	7.646×10^{-1}	m^3	ft^3 sec^{-1}	2.832×10^{-2}	m^3 sec^{-1}
fluid ounces	2.957×10^{-2}	liters, l *or* L	**Hydraulic conductivity**		
quarts	9.463×10^{-1}	l	U.S. gal day^{-1} ft^{-2}	4.720×10^{-7}	m sec^{-1}
U.S. gallons, gal	3.785	l	**Permeability**		
U.S. gal	3.785×10^{-3}	m^3	darcies	9.870×10^{-13}	m^2
acre-ft	1.234×10^{3}	m^3	**Tansmissivity**		
barrels (oil), bbl	1.589×10^{-1}	m^3	U.S. gal day^{-1} ft^{-1}	1.438×10^{-7}	m^2sec^{-1}
Weight, mass			U.S. gal min^{-1} ft^{-1}	2.072×10^{-5}	1 sec^{-1} m^{-1}
ounces avoirdupois, avdp	2.8349×10^{1}	grams, gr	**Magnetic field intensity**		
troy ounces, oz	3.1103×10^{1}	gr	gausses	1.0×10^{5}	gammas
pounds, lb	4.536×10^{-1}	kilograms, kg	**Energy, heat**		
long tons	1.016	metric tons, mt	British thermal units,	2.52×10^{-1}	calories, cal
short tons	9.078×10^{-1}	mt	BTU	1.0758×10^{2}	kilogram-meters, kgm
oz mt^{-1}	3.43×10^{1}	parts per million, ppm	BTU lb^{-1}	5.56×10^{-1}	cal kg^{-1}
Velocity			**Temperature**		
ft sec^{-1}(=ft/sec)	3.048×10^{-1}	m sec^{-1} (=m/sec)	°C + 273	1.0	°K (Kelvin)
mi hr^{-1}	1.6093	km hr^{-1}	°C + 17.78	1.8	°F (Farenheit)
mi hr^{-1}	4.470×10^{-1}	m sec^{-1}	°F - 32	5/9	°C (Celsius)

**Divide by* the factor number to reverse conversions.
Exponents: for example 4.047×10^{3} (see acres) = 4,047; 9.29×10^{-2} (see ft^2) = 0.0929.

Back Cover: Active mines and processing plants for industrial minerals in Wyoming, 1997, exclusive of sand and gravel operations. Compiled by the Wyoming State Geological Survey, 1997.



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