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# Geology of the Nevada Test Site and Nearby Areas, Southern Nevada

Scott Sinnock

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# Geology of the Nevada Test Site and Nearby Areas, Southern Nevada

Scott Sinnock
NTS Waste Management Overview Division 9764
Sandia National Laboratories
Albuquerque, NM 87185

## Abstract

The Department of Energy's Nevada Test Site (NTS) lies in the southern part of the Great Basin Section of the Basin and Range Physiographic Province. The NTS is characterized by alluvium-filled, topographically closed valleys surrounded by ranges composed of Paleozoic sedimentary rocks and Tertiary volcanic tuffs and lavas. The Paleozoic rocks are a miogeosynclinal sequence of about 13,000 ft of pre-Cambrian to Cambrian clastic deposits (predominantly quartzites) overlain by about 14,000 ft of Cambrian through Devonian carbonates, 8,000 ft of Mississippian argillites and quartzites, and 3,000 ft of Pennsylvanian to Permian limestones. During the Mesozoic an apparent depositional hiatus occurred. Tertiary volcanic rocks are predominantly silicic composition and were extruded from numerous eruptive centers during Miocene and Pliocene epochs. Within eruptive caldera depressions, volcanic deposits accumulated to perhaps 10,000 ft in total thickness, thinning to extinction outward from the calderas. Extrusion of minor amounts of basalts accompanied Pliocene and Pleistocene filling of structural basins with detritus from the ranges.

Regional compressional and extensional structures as well as local volcanic structures occur in the NTS region. Granitic intrusion accompanied compressional thrust faulting and folding of Paleozoic sedimentary rocks during regional Mesozoic mountain building. Normal extensional faulting coincided with the outbreak of volcanism during the Miocene and was superimposed on existing Mesozoic structures. Continued extensional deformation may be occurring at the present time. Strike-slip faulting and bending along the northwest trending Las Vegas Valley, Amargosa Desert, and Walker Lane produced as much as 30 to 50 mi of mid Tertiary, right-lateral movement in the western Great Basin. Currently local stresses apparently are being released along a series of northwest trending, left-lateral shear zones in the southern portion of the NTS.

# Acknowledgments

This report provides a partial synthesis of the geologic information relating to the Nevada Test Site and nearby areas in southern Nye County, Nevada. Though certain geological terms are used without adequate definition for non-geologists, definitions of many terms used in the text appear in the Dictionary of Geological Terms (American Geological Institute, 1976).

Because little of the material contained in this report is derived from my own field study, I am deeply indebted to previous workers who spent many years gathering geological field information about the NTS and its environs. Many conversations with personnel from the U.S. Geological Survey have been very helpful in catalyzing this report. Special thanks are extended to R. B. Scott of the U.S. Geological Survey for his thorough critique of a draft manuscript. Judy Salas of the Department of Geology, University of New Mexico, deserves special thanks for drafting most of the figures. If I have misrepresented the observations or ideas of previous workers, the responsibility is solely mine.

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# Geology of the Nevada Test Site and Nearby Areas, Southern Nevada

## Background and Objective

The Nevada Test Site (NTS) is federally owned, restricted access land located in southern Nye County, Nevada (Figure 1). It was selected in December 1950, as the Atomic Energy Commission's (1945 to 1974 predecessor agency of the Department of Energy) on-continent proving ground for nuclear weapons developmental testing. Originally, 640 sq mi* were withdrawn from the Air Force Las Vegas Bombing and Gunnery Range (presently Nellis Air Force Range). Subsequently, additional land withdrawals to the west (1954) and north (1964) increased the total area under restricted access to the present 1350 sq mi.

Activities conducted on the NTS are administered from Las Vegas, Nevada, by the Nevada Operations Office of the Department of Energy (DOE). Three national laboratories, Lawrence Livermore National Laboratory, Los Alamos National Laboratory and Sandia National Laboratories, propose and conduct weapon and related tests for the DOE on the NTS. Architectural, engineering, and construction support are provided by Reynolds Electrical and Engineering Company; Holmes and Narver, Incorporated; and Fenix and Scisson, Incorporated, among others. In 1950 the U.S. Geological Survey (USGS) was asked by the Atomic Energy Commission (AEC) to recommend which valleys on the Las Vegas Bombing and Gunnery Range were suitable for explosive cratering experiments.

The survey recommended Frenchman and Yucca Flats. Since that time the USGS has continued to evaluate the geological environment of the region in conjunction with long-range Great Basin studies. Other organizations and agencies including John A. Blume and Associates (seismic analysis), and the Environmental Protection Agency (radiation effects research), provide specialized information about the NTS and the effects of DOE's activities.

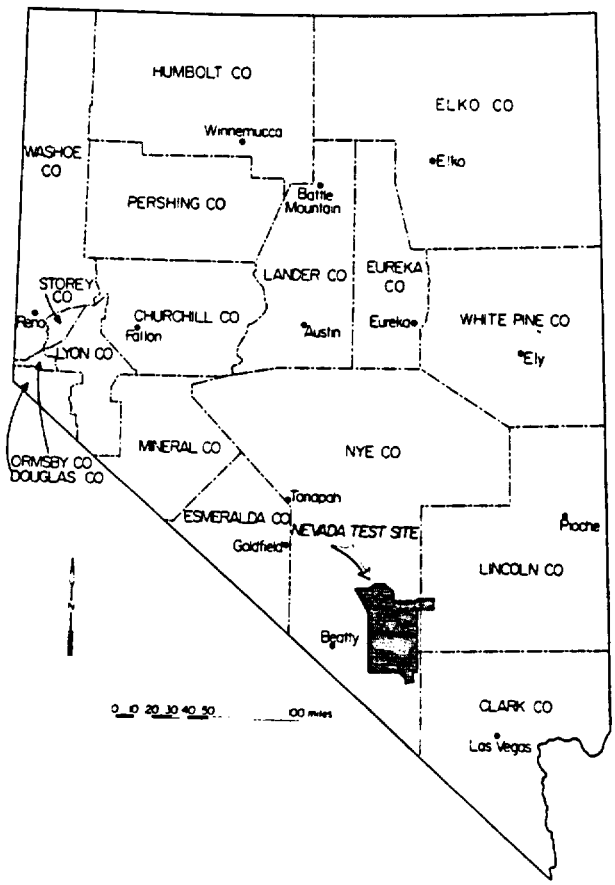


**Figure 1.** Location of the NTS in Nevada.

*English units of measurement are used throughout this report because the source information from which this report was generated was so expressed. The present report is a synthesis of previous studies and translation of units of measurement is, in the author's view, inappropriate.

Thus, a diversity of scientific, engineering, and logistical expertise is focused at the NTS. This is coupled with an existing management structure for administering withdrawn, controlled access land. For these reasons, the AEC offered the NTS for consideration as a permanent disposal site for high-level radioactive waste. In July 1972, the USGS reconnoitered the NTS for possible sites for the underground disposal of radioactive wastes. The results concluded that satisfactory sites might be found in underground masses of argillite and tuff (Jenkins and Thordarson, letter report). National policy at that time was focused on salt as a host rock for underground repositories for radioactive wastes (National Academy of Sciences, 1957; Bradshaw and McClain, 1971). This policy prevailed through the early 1970s, when recognition emerged that alternatives to safe disposal in salt were available (Schneider et al., 1974). An outgrowth of this recognition was the Energy Research and Development Administration's (1974-1974 predecessor agency to the Department of Energy) support of geological research to determine the best potential sites for waste disposal within the continental United States regardless of rock type (ERDA, 1976a, 1976b). Subsequently, ERDA selected the Nevada Operations Office to administer a program for evaluating the suitability of the NTS for radioactive waste disposal.

Early nuclear waste research at the NTS and elsewhere was directed primarily toward determining the response of potential host media to emplacement of heat-generating waste (Weaver, 1976; Parsons et al., 1976; Westinghouse, 1976). Continued policy evolution led to the current appreciation of the total geological environment in ensuring permanent waste isolation (de Marsily et al., 1977 and Bredehoeft et al., 1978). This report addresses the geological setting of the NTS in the context of the current waste isolation policy. The intent is to provide a synthesis of geological conditions at the NTS and nearby areas so that a general background of information is available for assessing the possible role of geology in providing protections for humans from buried radioactive wastes.

Hydrological conditions at the NTS and surrounding areas are summarized by Winograd and Thordarson (1975) and investigators from the Desert Research Institute of the University of Nevada (Fenske and Carnahan, 1975; Carnahan and Fenske, 1975; and Mifflin, 1968). These reports and their references provide a synopsis of hydrologic information similar to the geologic summary of this report. The "Summary" section of Winograd and Thordarson (1975) is appended in its entirety to this report to provide a brief, if somewhat technical, outline of hydrologic conditions in the NTS area.

# Physiography

## Topography

The NTS is located in the southern part of the Great Basin Section of the Basin and Range Physiographic Province (Fenneman, 1931; Hunt, 1974). The NTS and nearby areas occur on a broad topographic slope that separates high, topographically closed basins of central Nevada from low, connected basins of the Amargosa Desert-Death Valley drainage system. Topographically high areas of the NTS are dominated by a large volcanic plateau, Pahute Mesa, bounded on the north by the Belted and Kawich ranges and on the east by the Eleana Range (Figure 2). Pahute Mesa is flanked to the south and west by roughly circular, extinct volcanoes at Timber Mountain and Black Mountain. South of Timber Mountain, volcanic rocks similar to those on Pahute Mesa are broken by a series of faults into low, north-trending ridges and valleys of Yucca Mountain. Rainier Mesa and a southward-extending ridge form the eastern edge of Pahute Mesa, separating Timber Mountain from Yucca Flat. Southeast of Timber Mountain the Calico Hills form a structural and topographic dome with Paleozoic rocks exposed at its center. This dome forms the northern edge of an alluvial valley, Jackass Flats.

The eastern and southern portions of the NTS are characterized by a large basin dotted with small, irregularly oriented sub-basins and ranges. The overall basin contains Yucca, Frenchman, and Jackass Flats, and Mid, Emigrant, and Rock Valleys. The small ranges include, but are not limited to, Papoose Range, Halfpint Range, Buried Hills, Massachusetts Mountain, Lookout Peak, Mine Mountain, Syncline Ridge, C. P. Ridge, Ranger Mountain, Skull Mountain, Little Skull Mountain, and the Striped Hills. The disheveled nature of these small ranges obscures the overall basin character of the region. Therefore an observer is impressed with the apparent enclosure of each small basin by an unbroken skyline of continuous mountains. However, peaks within the ranges rarely exceed 6000 ft in elevation, and range crests generally rise only 500-1500 ft above the basin floor. The basin floor slopes from about 5000 ft elevation in the north to about 3000 ft in the south, where it merges with the Amargosa Desert. In contrast, the Belted Range-Pahute Mesa-Timber Mountain uplands exceed 7000-8000 ft in elevation and rise more than 2000 ft above the surrounding basins.

Alluvial fans characteristically occur at the bases of the ranges. These fans are generally featureless, gently concave slopes descending toward a central basin drainage line or playa (Figure 3).

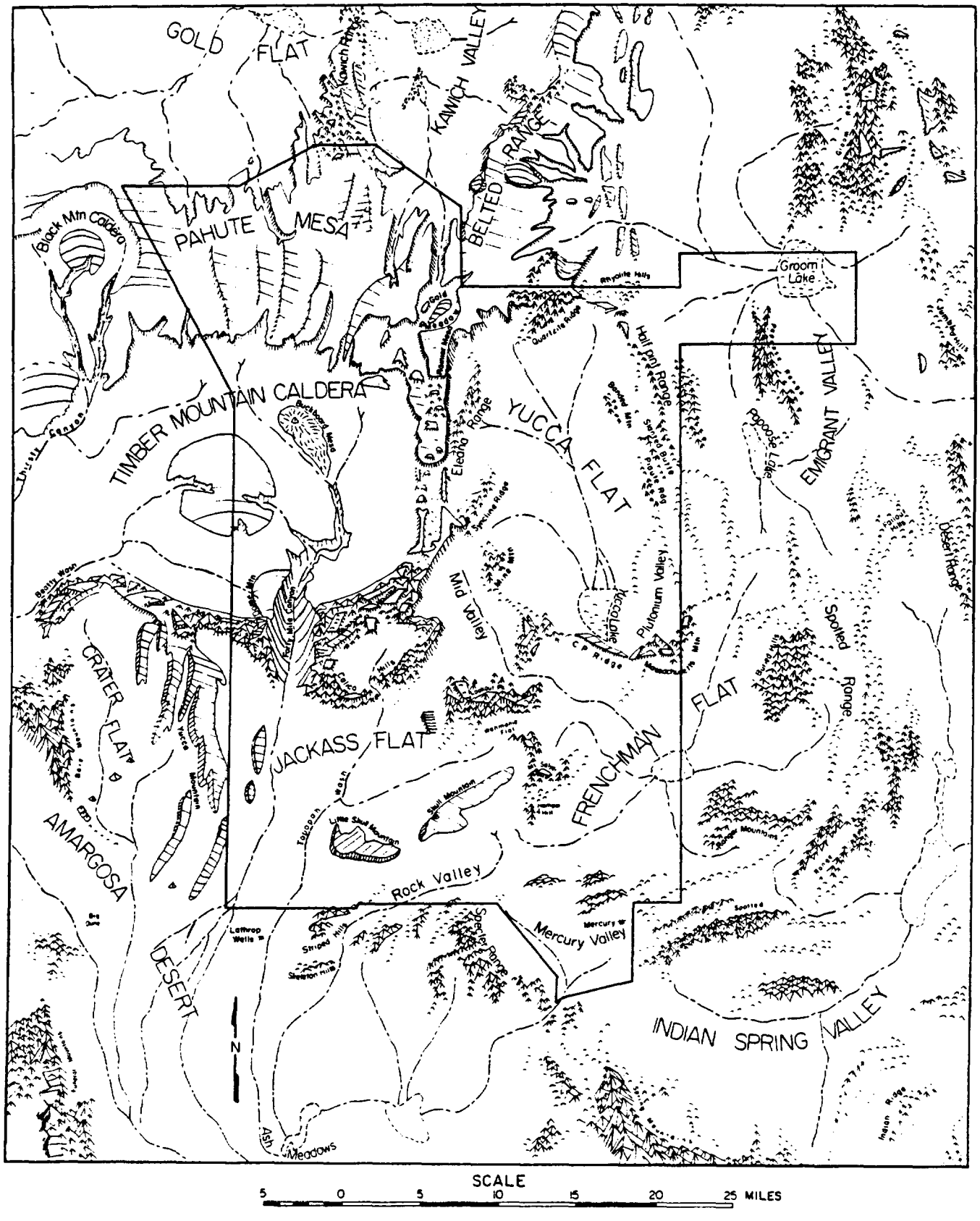


**Figure 2.** Physiographic Diagram of the NTS and Surrounding Areas

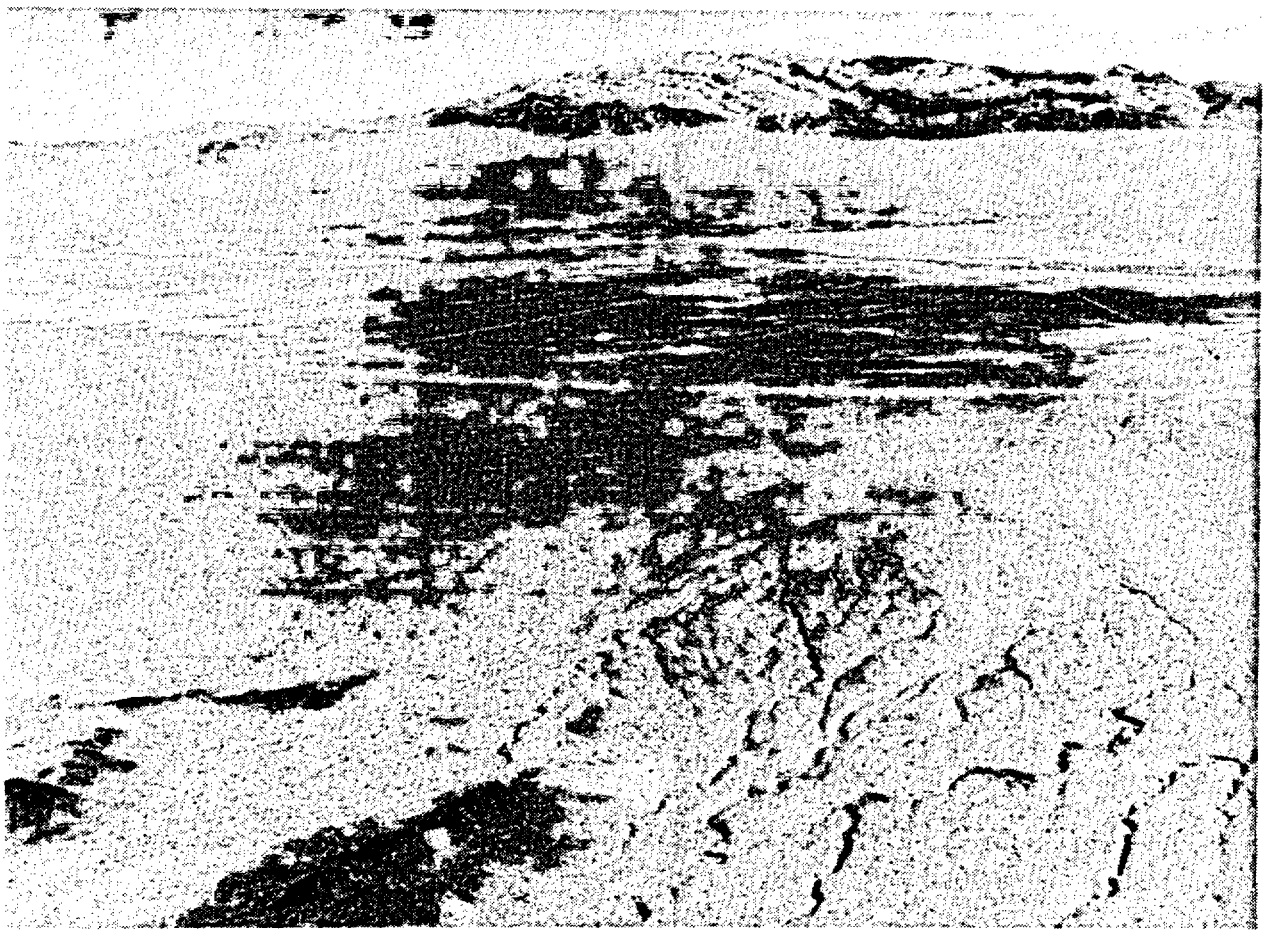

**Figure 3.** Characteristic Alluvial Valley in the Basin and Range. Dissected alluvial fans (foreground) from the Spring Mountain stream toward the center of Las Vegas Valley where they merge with coalescing fans at the base of the Sheep Range (background). Oblique aerial view is northeast from a point about 40 miles southwest of the NTS. Photo courtesty of Pan American, Inc.

Volcanic ranges at Pahute Mesa, Yucca Mountain and Skull Mountain are generally formed by relatively flat-lying rocks which exhibit a smooth skyline (Figure 4). Mountain slopes are generally deeply carved by widely spaced canyons. Non-volcanic ranges, in contrast, are generally formed by moderately to steeply tilted layers of Paleozoic sedimentary rocks and possess irregular skylines sharply punctuated by numerous peaks and mountain passes (Figure 5). Peaks commonly correspond to rocks which are relatively resistant to erosion such as quartzite and limestone, whereas the passes generally form upon less resistant shales and argillites.

## Drainage

Surface drainage on the NTS flows in eight discrete systems. Seven are locally closed and terminate in playa lakes; the eighth is the Amargosa River system which eventually flows to the Death Valley drainage sink. The northern portion of Pahute Mesa drains along deeply entrenched canyons to playas in Gold Flat and Kawich Valley. The southern flank of Pahute Mesa is similarly carved by the headwaters of Thirsty Canyon and Fortymile Canyon. Main branches of these streams as well as the headwaters of Beatty Wash flow through the Timber Mountain caldera and discharge to the Amargosa River. Fortymile Canyon is a spectacular anomaly in Great Basin topography, a water-carved gorge incised 2500 ft through the southern rim of Timber Mountain Caldera.

Jackass Flats, Calico Hills, Skull Mountain, and Little Skull Mountain drain to the Amargosa Desert via Topopah Wash and Rock Valley. Streams from the Specter Range and Striped Hills flow through Rock Valley and Ash Meadows to the Amargosa drainage. The eastern portions of the NTS drain to small playas at Groom Lake, Papoose Lake, Yucca Lake, and Frenchman Flat. All these playas contain lacustrine deposits and were formerly the sites of shallow Pleistocene lakes.

**Figure 4.** Typical Volcanic Range Topography at Yucca Mountain. Volcanic tuff units form smooth upper surface and banding along the side of the mountain. Note separate fault blacks, especially in the upper right portion of the photo. Oblique Aerial View is to the Southeast. Photo courtesy of Pan American, Inc.

**Figure 5.** Typical Character of Southern Great Basin Ranges Formed by Paleozoic Sedimentary Rocks. Note tilted bedding layers of the Spring Mountains in the foreground and irregular skylines of the Pintwater and Desert ranges in the background. Oblique aerial view is northeast toward the Indian Springs Valley playa. Photo courtesy of Pan American, Inc.

Streams on the NTS in particular and the Great Basin in general are ephemeral and flow only for brief periods during local or regional rains. The water table is deep below the surface, and runoff is quickly absorbed by gravels in the stream beds as the water flows beyond the area of local rainfall. Occasionally sufficient rain occurs, particularly from December through March, and the playas are covered with shallow, temporary lakes.

The ephemeral drainage is caused by scant rainfall and intense evaporation in the southern Nevada desert. Annual precipitation varies with elevation from about four inches in the basins to about ten inches on Pahute Mesa (Quiring, 1965). Meyers (1962) estimates about seventy inches of annual potential evaporation from the basins of southern Nevada. Hunt et al. (1966) measured a rate of 155.05 inches of pan evaporation per year in Death Valley, a hotter area than the NTS but exposed to a similar, harsh climatic environment. The large annual water-budget deficit causes all, or nearly all, precipitation to return to the atmosphere rather than to infiltrate to the regional water table. Beneath ephemeral washes and at higher elevations small amounts of rainfall may seep into the deep groundwater. Streams in the Paleozoic ranges are generally more closely spaced and less deeply entrenched than those in the volcanic ranges. In caldera complexes drainage is generally widely spaced and radial from a resurgent dome, annular in the moat, and centripital on the outer scarp.

# Rock Types

## General Statement

Within the stratigraphic record of the NTS region are preserved records of ancient Nevadan seas, mountain ranges, erupting volcanoes, and many other phenomena now ravaged by the geological forces of change. Those concerned with radioactive waste management need to recognize the ephemeral nature of any landscape in geological time. Though evolutionary processes have created and destroyed entire continental landmasses many times during the course of earth history, the rate of change is very slow from the perspective of a single human lifetime or even the perspective of all human history. The vulnerability to major changes of present geological conditions over the next few million years is very low, even in a generally active region such as southern Nevada. The time required for geological processes and individual events to occur should be borne in mind as the following stratigraphic evidence is reviewed.

Throughout the remainder of the report, geological time terms will be used repeatedly. Figure 6 summarizes the relations among these terms. Figure 7 is a stratigraphic column of the rocks in the NTS area. This column is included to provide a graphic reference for the following stratigraphic descriptions.

Rocks occurring at the NTS may be conveniently separated into three distinct groups: pre-Cambrian through Paleozoic sedimentary rocks; Tertiary volcanic deposits; and Tertiary through Recent basalts and unconsolidated sediments. In keeping with geological tradition, these rocks will be discussed beginning with the oldest.

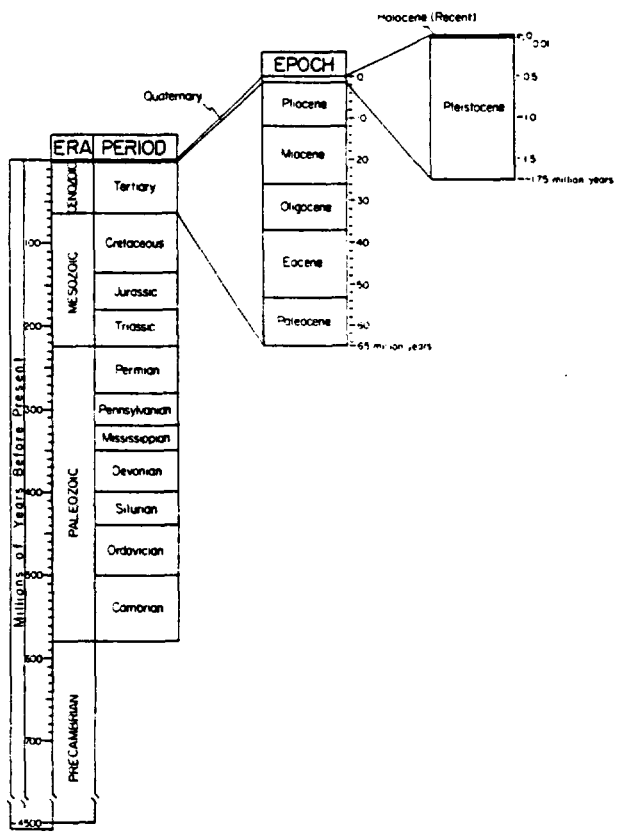


**Figure 6.** Geologic Time Scale (Compiled from Various Sources). Some interpretations of the absolute ages for epoch boundaries in the Tertiary differ from those shown on this chart

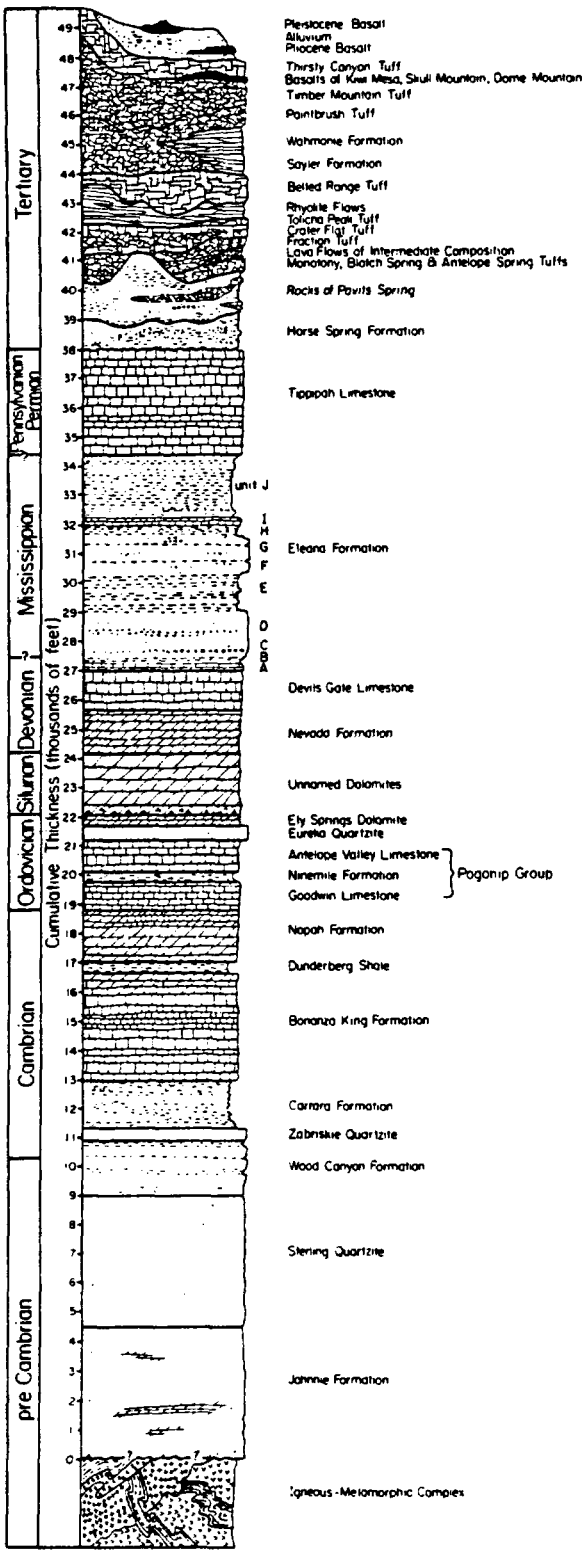


**Figure 7.** Idealized Stratigraphic Column of the NTS Area; Not All Rock Units are Present in All Portions of the Area, Especially Volcanic Units; Thickness of Units Likewise Varies Throughout the Area (various sources, see text)

## Paleozoic and Older Sedimentary Rocks

Pre-Cambrian through Permian rocks at the NTS represent a typical cordilleran miogeosynclinal sequence. Miogeosynclinal rocks are deposited in slowly subsiding depositional troughs on shallowly submerged edges of continental platforms (Figure 8). Deposits in miogeosynclines are generally derived from erosion of stable continental landmasses. A modern analogy for the Paleozoic Miogeosyncline of Nevada is the continental slope off the gulf coast of the United States. Miogeosynclines are commonly bounded seaward by eugeosynclines, deep-water troughs filled with sediments derived from volcanic island mountain chains, such as the present day Japanese and Phillipine Islands. Though eugeosynclinal rocks are not exposed in the NTS area, a Paleozoic island arc probably existed west of the ancient southern United States landmass.

Subsidence rates of miogeosynclines generally approximate rates of sedimentation, allowing thick sequences of deposits (up to 60,000 ft or more), to accumulate in shallow water. In the western United States miogeosynclinal rocks are generally well-sorted, fine-grained clastic rocks interbedded with massive carbonate deposits. Specific rock types include sandstone, quartzite, siltstone, shale, limestone and dolomite. No major angular unconformities exist, though minor disconformities are common.

The Cordilleran Miogeosyncline extended from Mexico to Alaska and existed throughout most of Paleozoic time (Kay, 1951; Stewart and Poole, 1974) (Figure 9). Local warping caused the Paleozoic shoreline to migrate on and off (transgression and regression, respectively) the continental platform at different times in different places. Resulting sedimentary facies (Figure 10) and disconformities likewise differ locally along the extent of the trough. The general aspect of the miogeosynclinal deposits preserves a record of three and perhaps four major Paleozoic transgression-regressions (Sloss and Speed, 1974). The NTS lies where the seaward margin of the ancient miogeosyncline occurred, so this region did not directly experience most of the transgressions and regressions. Instead ancient shoreline migrations are recorded in the NTS area by vertical facies changes in an almost time-continuous, 40,000 ft thick sequence of marine deposits.

## Pre-Cambrian

The first transgression in southern Nevada is represented by the pre-Cambrian Johnnie Formation (Burchfiel, 1964). Its base is not exposed at the NTS,

but to the south it is locally underlain by the Noonday Dolomite (Wright et al., 1978), which, in turn, unconformably rests on a pre-Cambrian metamorphic and igneous shield complex (Cloud and Semikhatov, 1969; Williams et al., 1974). However, in the NTS region the Johnnie Formation is the basal sedimentary unit. It is predominantly quartzite, sandstone, and siltstone with minor interbeds of dolomite. Total thickness is about 4500 ft in the Specter Range just south of the NTS (Burchfiel, 1964).

The Johnnie Formation is conformably overlain by, successively, the pre-Cambrian Sterling Quartzite (Burchfiel, 1964; Ekren et al., 1971), and the pre-Cambrian to Lower Cambrian Wood Canyon Formation (Burchfiel, 1964; Stewart and Poole, 1974; Stewart, 1980). The Sterling is predominantly quartzite and ranges in thickness from 3700 ft in the Spring Mountains (Nolan, 1929) to almost 5300 ft at Quartzite Mountain (Ekren et al., 1971).

The Wood Canyon Formation is also predominantly quartzite, but it contains numerous beds of quartzitic siltstone and micaceous silty and sandy shale. An upward increase of finer-grained sediments in the Wood Canyon indicates an eastward transgression of the shoreline during the pre-Cambrian to Cambrian transition. Total thickness of the Wood Canyon is about 2300 ft in both the Specter Range (Burchfiel, 1964) and the Groom Lake area (Barnes and Christiansen, 1967). Ekren et al., (1971) report 3750 ft of Wood Canyon in the Belted Range, but suggest the anomalous thickness may be related to faulting and intrusion of Tertiary rhyolite dikes.

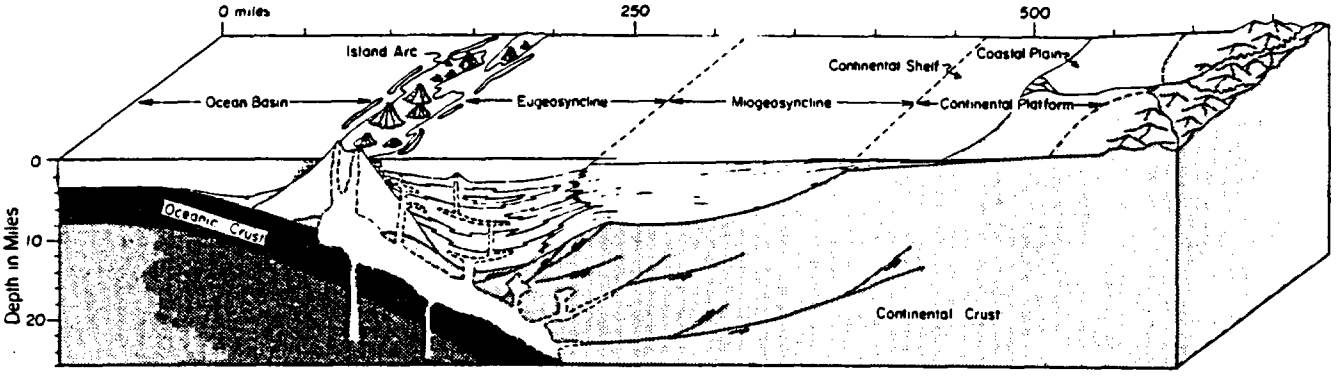


**Figure 8.** Idealized Features Along an Active Continental Margin, Including: Continental Platform, Miogeosyncline, Eugeosyncline, Volcanic Island Arc, and Deep Ocean Basins

**Figure 9.** Location of Paleozoic Cordilleran Miogeosyncline in the Western United States (after Stewart and Poole, 1974)

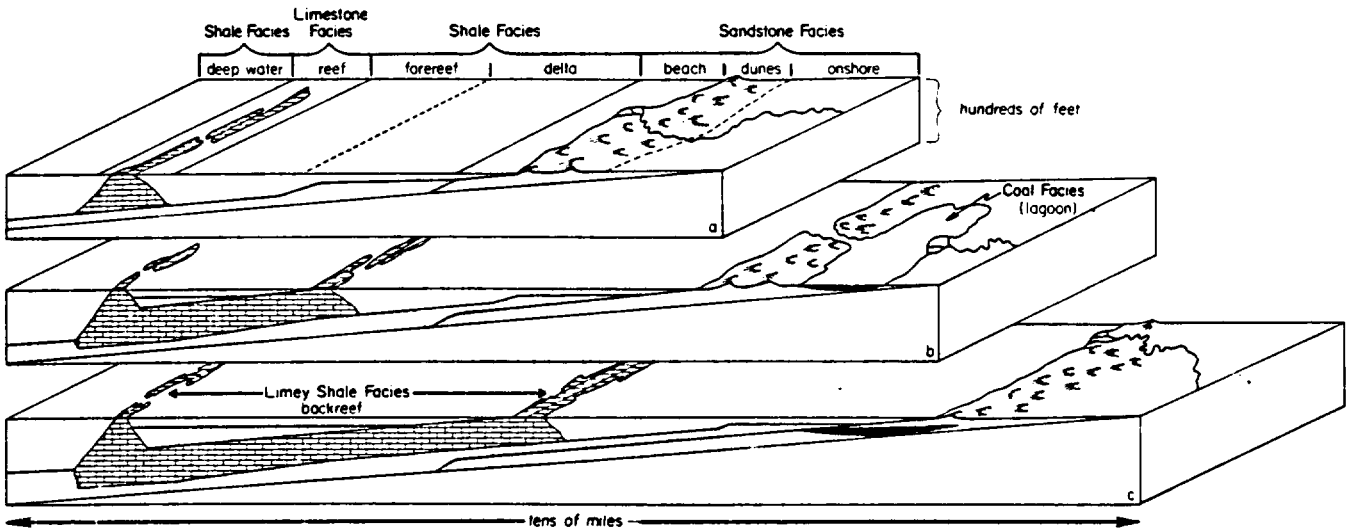


**Figure 10.** Idealized Lateral Migration of Transgressive Facies Over Time (Top to Bottom, Diagrams A to C) Showing the Resulting Development of Vertical Changes in Lithology

## Cambrian

The uppermost quartzite layer of the Wood Canyon Formation is considered by some authors as a separate formation, the Lower Cambrian Zabriskie Quartzite (Ekren et al., 1971; Barnes and Christiansen, 1967). This unit is only a few hundred feet thick near the NTS, but Cornwall and Kleinhampl (1961, 1964) report 1150 ft of time equivalent rocks, the Corkscrew Quartzite, at Bare Mountain. Deposition of the Zabriskie Quartzite culminated the accumulation of more than 10,000 ft of pre-Cambrian and Lower Cambrian quartzites and related rocks in the NTS area. Figure 11 shows the present distribution of outcrop areas of this massive quartzite assemblage.*

By middle Early Cambrian time, near shore, sandy deposits in the NTS region were succeeded by deeper water, offshore shale facies of the Carrara Formation. This clastic transition corresponds to transgression of the Tapeats Sandstone across the crystalline basement in the Grand Canyon district (McKee, 1945), the New York Mountains (Burchfiel and Davis, 1977), and the similar Flathead Sandstone transgression in western Montana (Dott and Batten, 1971). The Carrara Formation consists predominantly of about 1500 ft of shale, interbedded with minor siltstones and quartzites at the bottom and limestones at the top (Burchfiel, 1964; Ekren et al., 1971; Barnes et al., 1962). This formation marks the time of a transition between clastic and carbonate facies of the early Paleozoic miogeosyncline.

Overlying the Carrara is the Middle to Upper Cambrian Bonanza King Formation (Figure 12). The Bonanza King ranges from about 3000 ft thick in the Specter Range (Burchfiel, 1964) to 4600 ft in the Jangle Ridge-Banded Mountain area (Barnes et al., 1962). It is predominantly limestone with increasing amounts of dolomite toward the top. Cornwall and Kleinhampl (1964) report that the Bonanza King is absent west of NTS.

Resting on the Bonanza King is a 200-300 ft thick unit of shale with minor limestone layers. This unit, the Dunderburg Shale, is variously considered a formation (Burchfiel, 1964) or basal member of the overlying Nopah Formation (Ekren et al., 1971; Cornwall and Kleinhampl, 1961, 1964). It probably represents a minor pulse of uplift and increased erosion to the east causing a thin clastic blanket to spread across the miogeosynclinal trough.

Carbonate deposition resumed with the Upper Cambrian Nopah Formation (Burchfiel, 1964). The Nopah Formation predominantly consists of dolomite with a few limestone interbeds. Structural displacements make its thickness difficult to determine. Burchfiel (1964) reports about 1000 ft in the Specter Range; Barnes and Christiansen (1967), 2000 ft in the Groom District; and Ekren et al. (1971) suggest that perhaps 3000 ft exist in the Belted Range.

*Figure 11 and similar maps on Figures 13, 14, 16, 17, 18, 22 and 26 are somewhat unconventional geologic maps in that they show only the surface distribution of a single rock type or various lithologically similar rocks. This accentuates locations where particular rock types occur in accordance with radioactive waste disposal objectives to identify and characterize specific rock media suitable for hosting a mined nuclear waste disposal facility.

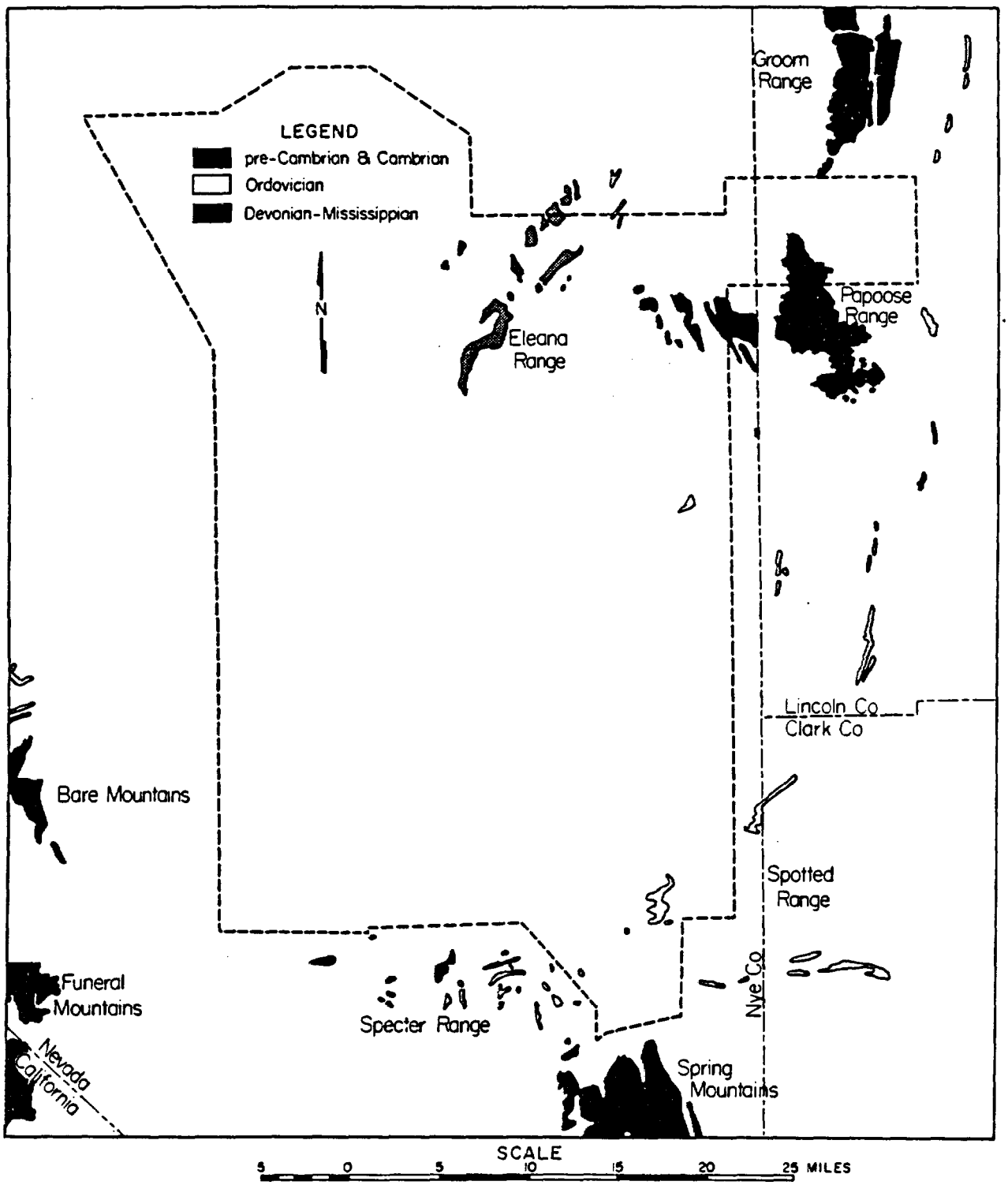


**Figure 11.** Distribution of Quartzite Outcrops in the NTS Area (from Cornwall, 1972, Tschanz and Pampeyan, 1970, Longwell, et al., 1965b, and USGS GQ map series of the NTS region)

**Figure 12.** Cambrian Bonanza King Formation Forms the Hills in the Background. Foreground shows vegetation associated with groundwater discharge at seeps and springs in Ash Meadows. Oblique aerial view is northeast toward the southwest portion of the NTS. Photo courtesy of Pan American, Inc.

## Ordovician

The Nopah Formation is conformably overlain by lower Ordovician carbonates assigned to the Pogonip Group. This group of rocks is composed of three formations in the NTS area; the Goodwin Limestone, the Ninemile Formation, and the Antelope Valley Limestone. The Goodwin Limestone is about 1000 ft thick and overlies dolomites of the Nopah Formation throughout the NTS area. (Ekren et al., 1971; Byers et al., 1961). Resting on the Goodwin Limestone is a thin (< 300 ft) layer of shale, the Ninemile Formation (Ross, 1964; Ekren et al., 1971; Byers et al., 1961; Burchfiel, 1964). This unit is similar to the Dunderburg Shale but contains more thin, interlaminated beds of siltstone and limestone. Considered together, the Carrara, Dunderburg, and Ninemile Formations make up about 2100 ft of lower Paleozoic argillaceous rocks. Their surface distribution in the NTS area is shown on Figure 13.

Carbonate deposition resumed with the Middle Ordovician Antelope Valley Limestone which contains increasing silt content and sandstone interbeds toward the top. It ranges from less than 1000 ft thick in the Specter Range (Burchfiel, 1964) to about 1600 ft in the Belted Range (Ekren et al., 1971).

Slower subsidence of the miogeosyncline is indicated by decreased thickness of the entire Pogonip Group to the west of the NTS. Cornwall and Kleinhampl (1961) and McAllister (1952) report a total thickness of about 1300 ft for the Pogonip Group at Bare Mountain and the Panamint Range of eastern California, respectively. Westward thinning of the Pogonip Group and increasing clastic content of the upper members forshadowed a Middle Ordovician regression that occurred throughout North America (Kay, 1951; Sloss and Speed, 1974).

In the NTS area deposition of the lower beds of the Eureka Quartzite accompanied this continental emergence (Ross, 1964; Webb, 1958). The Eureka conformably rests on the Antelope Valley Limestone. The regression during Eureka time marked the end of the first major Phanerozoic submergence of the continent, the Sauk submergence (Sloss, 1963). During upper Eureka time the sea readvanced across southern Nevada, reworked the lower Eureka sands, and left in its wake an upper transgressive Eureka sand deposit (Ketner, 1968). The upper Eureka Quartzite is the western equivalent of the well known transgressive St. Peters Sandstone of the eastern United States (Kay and Crawford, 1964) indicating the continent-wide extent of the Middle Ordovician transgression. The Eureka generally thickens from southeast to northwest approaching 400 ft in the central NTS area (Ross and Longwell, 1964), and 485 ft in the Specter Range (Burchfiel, 1964). This formation unit contains minor interbeds of siltstone, argillite and dolomite (Byers et al., 1961; Poole, 1965).

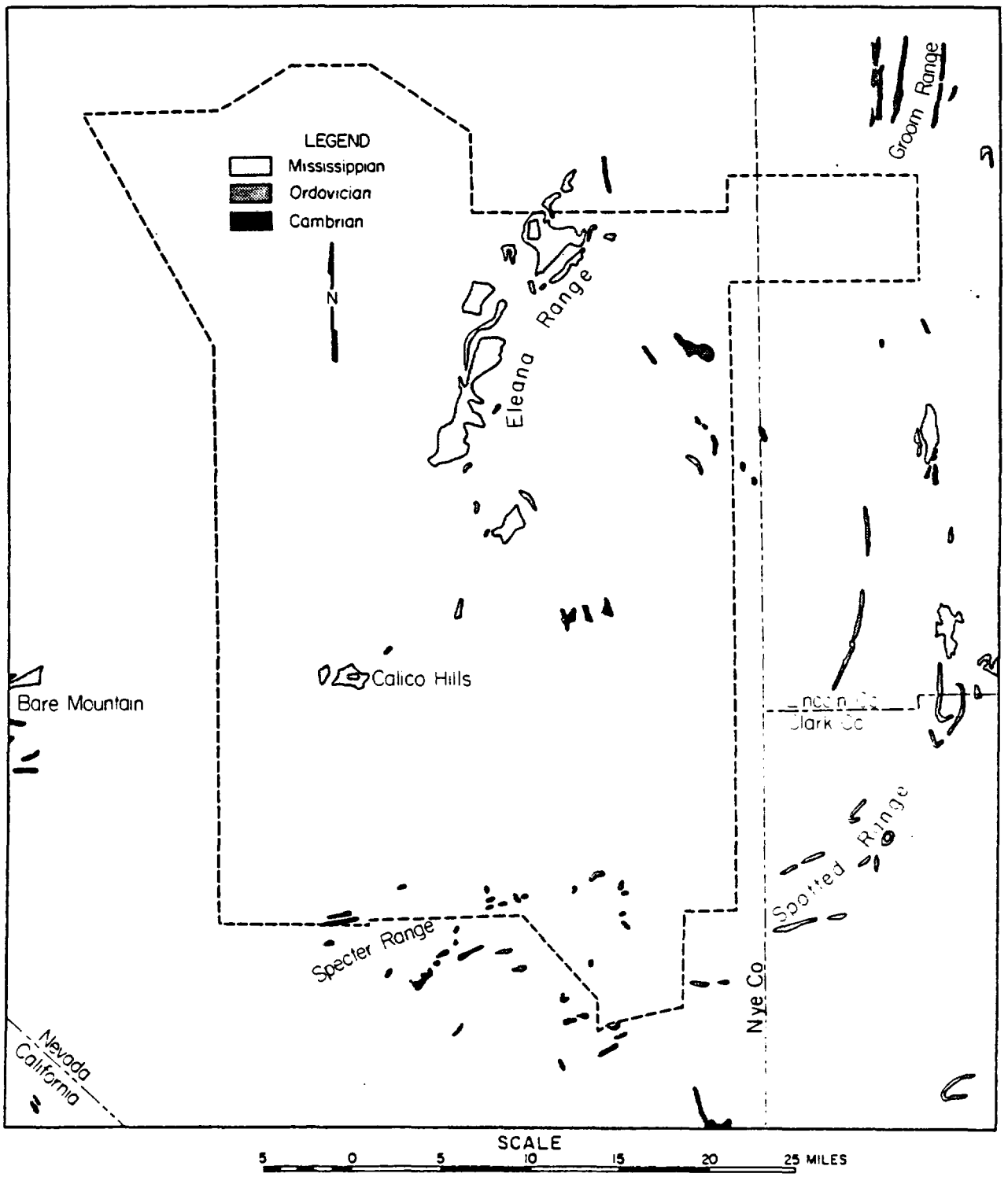


**Figure 13.** Distribution of Shale, Argillite and Some Siltstone Outcrops in the NTS Area (from Cornwall, 1972, Tschanz and Pampeyan, 1970, Longwell, et al., 1965b, and USGS GQ Map Series of the NTS Region).

The Eureka Quartzite is conformably overlain by the Upper Ordovician Ely Springs Dolomite. The Ely Springs contains quartz sand clasts in the lower part and considerable chert throughout (Burchfiel, 1964; Ekren et al., 1971). Its thickness is generally less than 300 ft in the NTS area. However, the total thickness of uninterrupted carbonate rocks above the Eureka Quartzite approaches 5000 ft. Besides the Ely Springs Dolomite, post-Eureka carbonates include about 2300 ft of locally unnamed Silurian dolomites, 1500 ft of the Lower and Middle Devonian Nevada Formation, and 1300 ft of the Upper Devonian Devils Gate Limestone (Burchfiel, 1964; Ekren et al., 1971; Johnson and Hibbard, 1957).

## Silurian

The Silurian dolomites rest conformably on the Ely Springs Dolomite. Burchfiel (1964) suggests they may correlate with the Silurian Roberts Mountain Formation and Lone Mountain Dolomite of the Eureka District (Nolan et al., 1956). Ekren et al. (1971) informally designate the Silurian section as the dolomite of the Spotted Range and suggest it may include rocks of Early Devonian age in its upper part. The Silurian dolomites are generally massive with chert interbeds in the lower portion and quartz sand disseminated in the uppermost beds. The Silurian section is the least studied group of rocks in the NTS area and future studies are required to substantiate its stratigraphic correlation.

## Devonian

The overlying Devonian Nevada Formation is also predominantly dolomite, reported as 1605 ft thick in the Specter Range (Burchfiel, 1964). About 1300 ft of the Devils Gate Limestone rests on the Nevada Formation. The Devils Gate contains abundant quartzite beds in the upper part (Johnson and Hibbard, 1957). The Devils Gate Limestone marks the upper unit of a massive carbonate section more than 12,000 ft thick. The surface distribution of this lower to middle Paleozoic carbonate sequence is shown in Figure 14.

## Mississippian

Although a pre-Late Devonian angular unconformity is variously expressed south and east of the NTS (Burchfiel and Davis, 1977; Sloss, 1972), only a possible minor disconformity occurs between the Devils Gate Limestone and the overlying Eleana Formation in the NTS area (Poole et al., 1961). The lowermost Eleana deposits are limestones and limestone-conglomerates assigned to the Late Devonian (Poole et al., 1965a). Poole et al., (1961) suggest that the limestone-conglomerates were derived from erosion of Devils Gate equivalents elevated by the earliest pulses of mountain building to the north. Johnson and Hibbard (1957) assign the lower Eleana carbonate beds to the Narrow Canyon and Mercury Limestone Formations. These Upper Devonian-Lower Mississippian carbonates are approximate correlatives of the lower Monte Cristo Limestone of the Spring and New York Mountains (Burchfiel and Davis, 1977; Bissell, 1974; Nolan, 1929).

The Eleana Formation grades upward to a thick clastic sequence of Mississippian argillite, quartzite, and conglomerate (Poole et al., 1961, 1965a). To the south (Burchfiel and Davis, 1977; McAllister, 1952; Nolan, 1929) and east (McKee, 1945, Dott and Batten, 1971) carbonate deposition continued throughout the Mississippian, whereas to the north, the clastic Chainman Shale and Diamond Springs Formation were deposited during Eleana time (Nolan et al., 1956; Roberts et al., 1958; Smith and Ketner, 1968; Bissell, 1974). Apparently, the NTS lies along a Late Devonian through Mississippian facies transition zone, separating clastic deposits derived from mountains in central Nevada from carbonate shelf deposits to the south and east (Poole, 1974). The Devonian-Mississippian Mountain building in central Nevada is known as the Antler Orogeny (Stewart, 1980).

Distinct units of the Eleana Formation are designated from the bottom to top as: A, carbonate; B, argillite; C, quartzite and conglomerate; D, argillite and quartzite; E, argillite; F and G, quartzite, argillite, and conglomerate; H, argillite; I, limestone and argillite; and J, argillite. Argillites (units B, E, H and J) account for about 5000 ft of the total formation thickness of about 7500-8500 ft. Quartzites (units C, D, F and G) comprise most of the remainder with limestones (units A and H) contributing only minor amounts. Outcrop areas of the argillite units of the Eleana Formation in the NTS area are shown on Figure 13. Quartzite units are included on Figure 11.

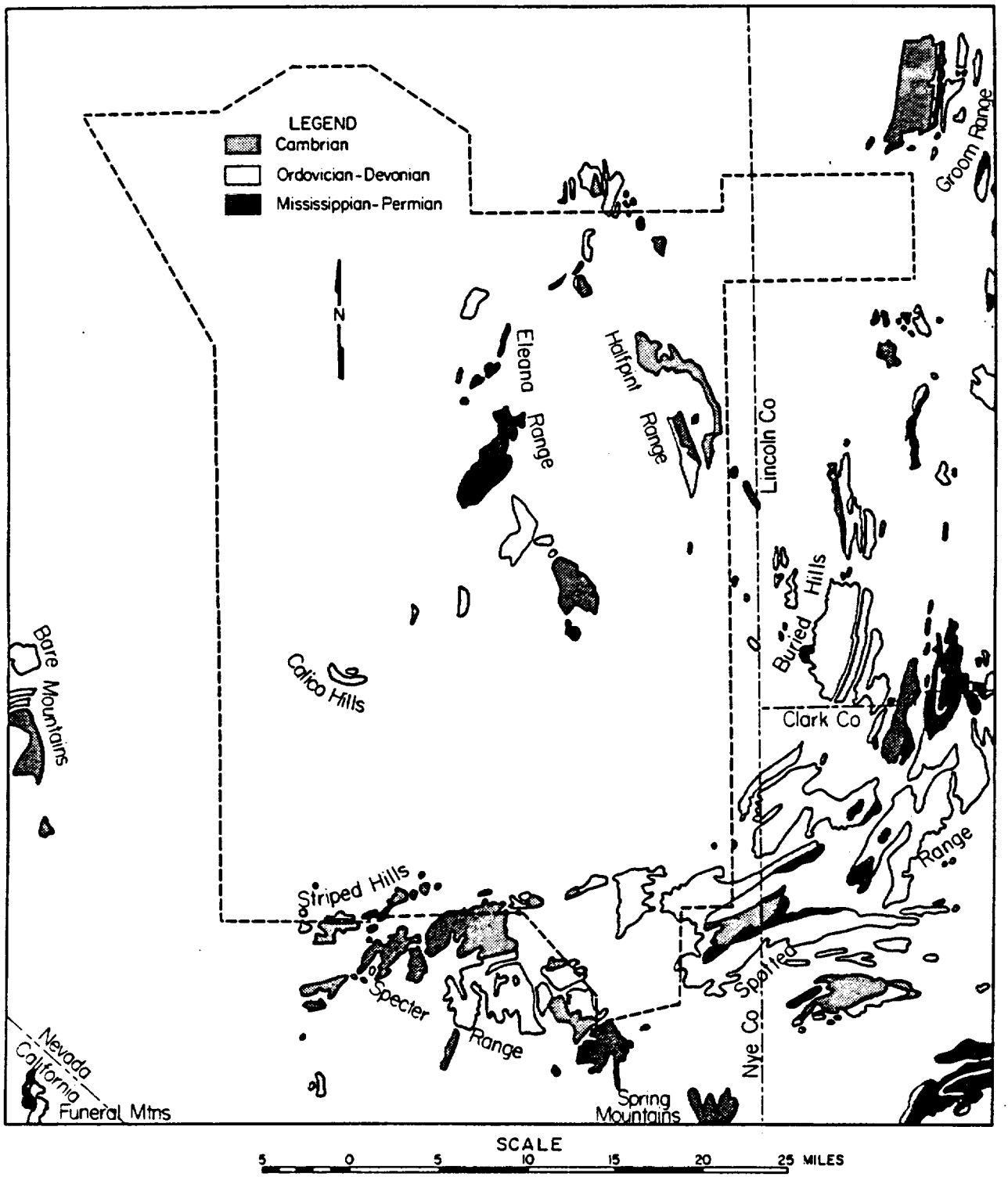


**Figure 14.** Distribution of Carbonate Outcrops in the NTS Area (from Cornwall, 1972, Tschanz and Pampeyan, 1970, Longwell, et al., 1965b, and USGS GQ map Series of the NTS Region)

## Pennsylvanian-Permian

About 3500 ft of Pennsylvanian and Permian Tippipah Limestone accumulated in the NTS area following an Early Pennsylvanian depositional hiatus (Gordon and Poole, 1968). The Tippipah Limestone is roughly equivalent to the Bird Springs formation of the Spring Mountains (Longwell and Dunbar, 1936; Rich, 1961; Burchfiel et al., 1974; Bissell, 1974) and New York Mountains (Burchfiel and Davis, 1977), and the Tihvipah Limestone near Death Valley (McAllister, 1952). The outcrop distribution of these upper Paleozoic carbonates is included with the lower to middle Paleozoic carbonates shown on Figure 14. The Permian Tippipah Limestone is the last evidence for marine submergence and related sedimentation in the NTS area. A highly generalized stratigraphic section of the Paleozoic deposits prior to subsequent deformation is provided in Figure 15.

## Mesozoic Intrusive Rocks

The Mesozoic Era is represented at the NTS only by widely scattered, small granitic intrusions; Climax, Twin Ridge, and Gold Meadows stocks. Gibbons et al. (1963) report an age of 130-140 m.y. for the Gold Meadows stock, placing the time of intrusion near the Jurassic-Cretaceous boundary. The Climax Stock has been dated as 93 m.y. (Maldonado, 1977; Naesser and Maldonado, 1981), or middle Cretaceous, though Barnes et al. (1963) assign it a Permian to Early Mesozoic age. Maldonado (1981) suggests the Twin Ridge stock may be genetically related to the Climax Stock. The distribution of intrusive outcrops in the NTS area is shown on Figure 16.

Triassic sedimentary shelf rocks exist to the south and northwest (Bissell, 1974) and Jurassic shelf rocks occur to the south, southwest (Burchfiel and Davis, 1971; 1977; Burchfiel et al., 1974) and southeast (Longwell, 1949). The Mesozoic was most likely a time of emergence, mountain building and erosion in the NTS area. Granitic intrusions in the area, a widespread angular unconformity between Paleozoic and Tertiary rocks, and the nearby occurrence of Mesozoic volcanic rocks (Burchfiel and Davis, 1977) support this interpretation. Mesozoic uplifts in Nevada have been variously referred to (from oldest to youngest) as the Sonoman Orogeny (Armstrong, 1968a), Nevadan Orogeny (Roberts, et al., 1958), Sevier arch (Armstrong, 1968b), Sevier Orogeny (Harris, 1959), and Laramide Orogeny (Longwell, 1952). Regardless of the exact timing or sequence of separate uplifts, the Mesozoic Era produced the final destruction of the depositional environment of the Cordilleran Miogeosyncline.

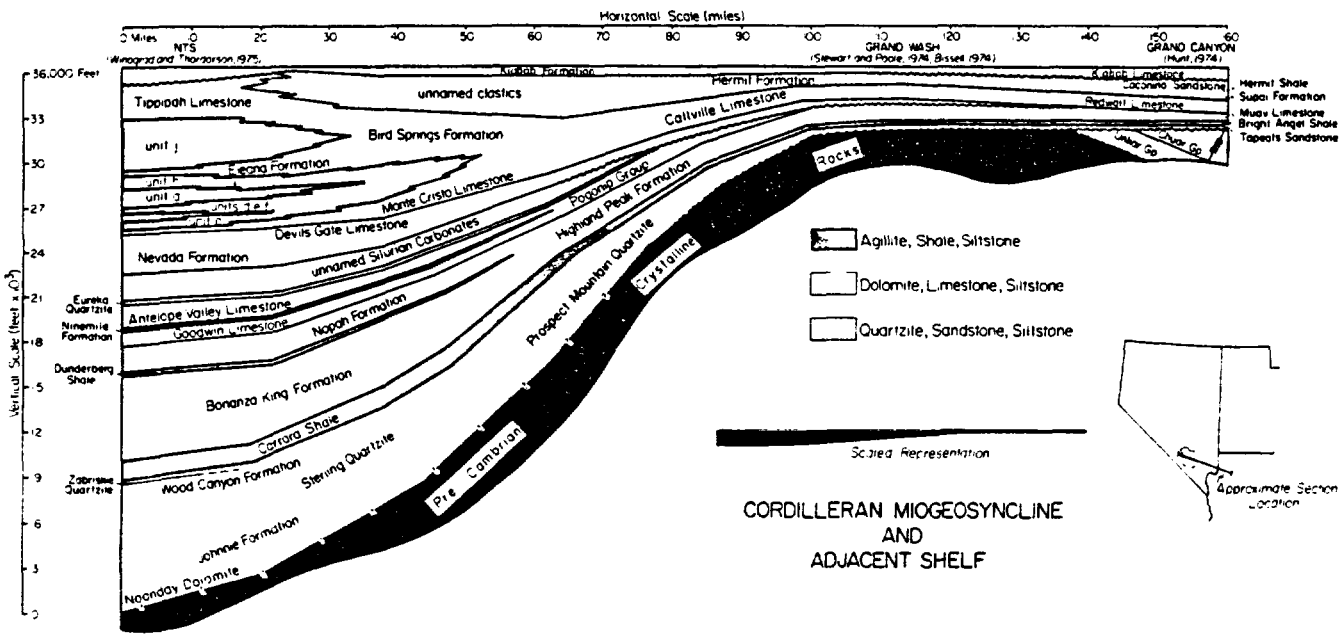


**Figure 15.** Generalized Stratigraphic Section of the Cordilleran Miogeosyncline and Adjacent Shelf in Southern Nevada and Northern Arizona Prior to Subsequent Structural Deformation

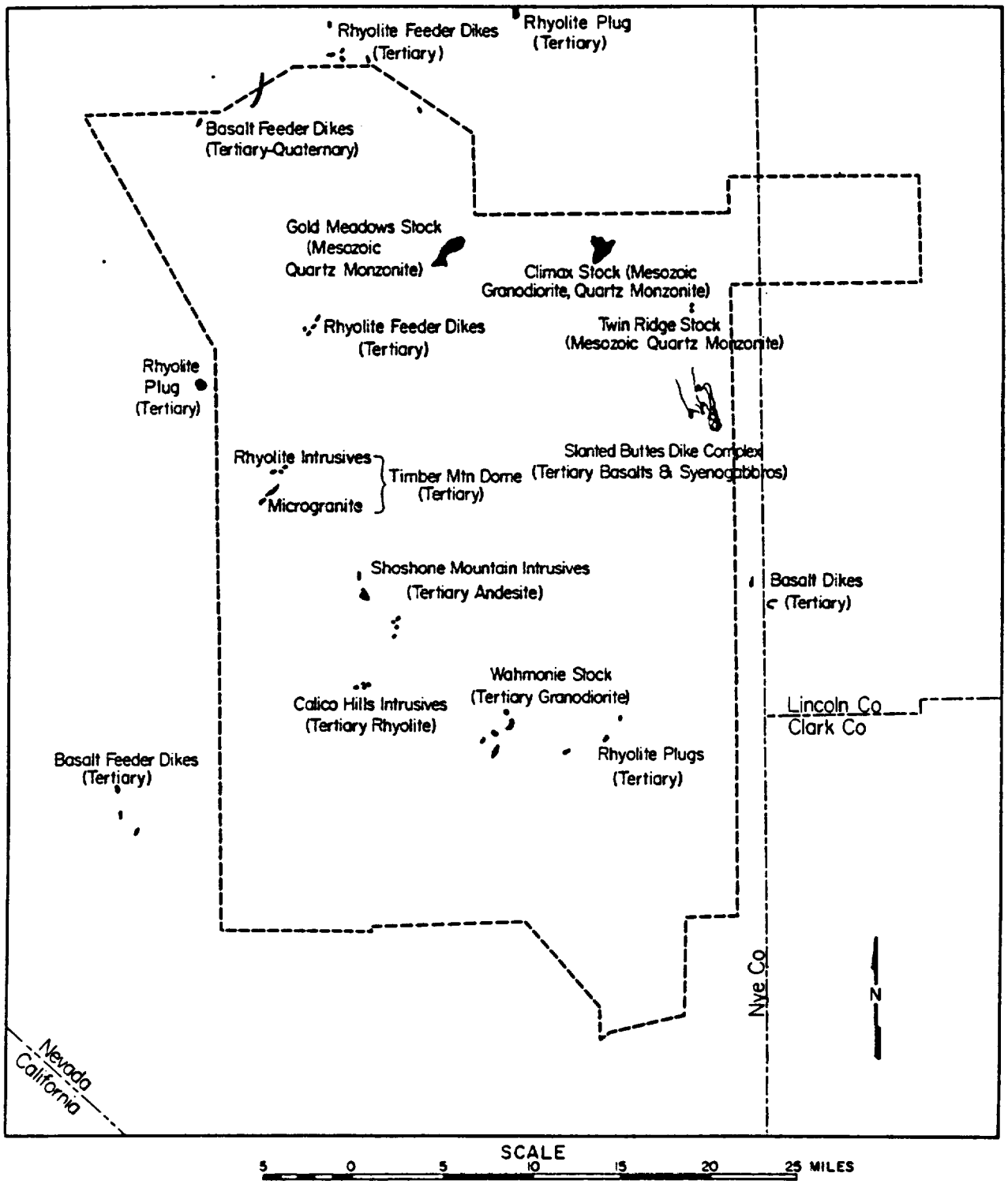


**Figure 16.** Distribution of Intrusive Outcrops in the NTS Area (from Cornwall, 1972, Tschanz and Pampeyan, 1970, Longwell, et al., 1965b, and USGS GQ Map Series of the NTS Region)

## Cenozoic Volcanic Rocks

Brief tectonic quiescence apparently followed the Mesozoic mountain building in southern Nevada. By late Oligocene time (~25-30 m.y. ago), erosion had reduced the ancient ranges to a low to moderate relief plain. A thin layer of silts, clays, and limey muds veneered this plain, perhaps as pediment, flood-plain, and local lacustrine deposits. Upon this surface great volumes of Miocene and Pliocene volcanic materials were deposited.

### Pre-Volcanic Sedimentary Rocks

In the NTS region, the sedimentary veneer is variously named the Horse Springs Formation, Rocks of Mara Wash, and Pavits Springs Rocks and is composed of interbedded siltstone, claystone, conglomerate, limestone, tuffaceous sandstone, and minor tuff (Hinrichs, 1968a). To the west, equivalent rocks compose the Titus Canyon Formation (Cornwall and Kleinhempl, 1964; Stock and Bode, 1935). Tuff deposits are abundant in the Rocks of Pavits Spring which generally overlie the Horse Spring Formation. An upward increase in tuff and tuffaceous sandstone signals the outbreak of volcanic activity near the NTS area.

Earlier silicic volcanism had blanketed northern Nevada beginning about 40-45 m.y. ago (Snyder et al., 1976; Armstrong, 1970; Stewart et al., 1977). However, the first volcanic activity near the NTS apparently commenced during Horse Springs time. Hinrichs (1968a) reports an age of 29 m.y. for a tuff in the Horse Springs Formation in the NTS area. Armstrong (1970) dated the age of the Horse Springs Formation at 21.3 m.y.

A thick section of predominantly volcanic tuff accumulated in the NTS area after Horse Spring time. In 1957, Johnson and Hibbard (1957) classified all Tertiary volcanic rocks comprising this section as the Oak Springs Formation. Later work resulted in recognition of eight distinct members originating from volcanoes on or near the NTS (Hinrichs and Orkild, 1961). Poole and McKeown (1962) divided the same rocks into a lower Indian Trails Formation and upper Piapi Canyon Formation. These two units were further subdivided into six formations in 1964, from oldest to youngest: 1) Indian Trails Formation; 2) Belted Range Tuff; 3) Salyer Formation; 4) Wahmonie Formation; 5) Paintbrush Tuff; and 6) Timber Mountain Tuff. Each formation consists of various members (Sargent et al., 1965; Poole et al., 1965b; Orkild, 1965). This terminology generally has persisted to the present with deletion of the Indian Trails Formation and elevation of the Stockade Wash Tuff and Red Rock Valley Tuff to formation status (Byers et al., 1976a,b).

Older volcanic units in the NTS area were extruded from volcanoes some distance away from the NTS. These units include the Monotony, Fraction, Tolicha Peak, and Kane Springs Wash Tuffs. Other volcanic units in areas peripheral to and extending into the NTS are also generally recognized; the Crater Flat and Thirsty Canyon Tuffs, and the rhyolites of Calico Hills and Fortymile Canyon (Christiansen et al., 1977).

Late Tertiary volcanic rocks at the NTS are overwhelmingly composed of air-fall and ash-flow tuff (Figure 17) except the Salyer and Wahmonie Formations and the rhyolites of Calico Hills, Fortymile Canyon and Shoshone Mountain which contain abundant silicic lava flows (Figure 18). Minor interbeds of basalt, andesite, rhyolite, tuffaceous alluvium, breccia flows, and a myriad of other rock types of volcanic provenance occur within the Tertiary section at the NTS. The Belted Range Tuff and Thirsty Canyon Tuff are peralkaline (Sargent et al., 1965; Byers et al., 1976a); the remainder are generally silicic. Table 1 summarizes the history of volcanic events and resulting deposits in the NTS area.

Both ash-flow and air-fall varieties of tuff are common around silicic eruptive centers. Ash-flow deposits are formed by the settling, compaction, and fusion of hot gas and lava clouds (nueé-ardentes) which race down volcanic slopes during explosive eruptions. Commonly, they form a single cooling unit characterized by a densely welded central portion surrounded above, below, and distal from the eruptive source by less welded parts. A glassy unit, or vitrophyre, often is found at the base of ash-flow units. Air-fall tuffs are formed by ash which is ejected into the atmosphere, cooled, and deposited as a blanket downwind from the eruptive source. Air-fall tuffs are generaly nonwelded, porous, low density rocks in contrast to the more densely welded, ash-flow units. If multiple ash-flow units are extruded within a relatively short time (days to weeks) complete cooling of the earlier flows will not have had time to occur, and compound cooling units will form. These compound cooling units may be characterized by complex patterns of welding including moderately welded zones between distinct ash-flows and perhaps by partially welded air-fall units sandwiched between densely welded ash-flows. Thickness of individual units varies depending upon the paleotopography of the depositional surface and distance from the source vent. Ash deposits generally thin away from eruptive centers. More detailed discussion of the physical properties of silicic tuff deposits is available in Smith (1960).

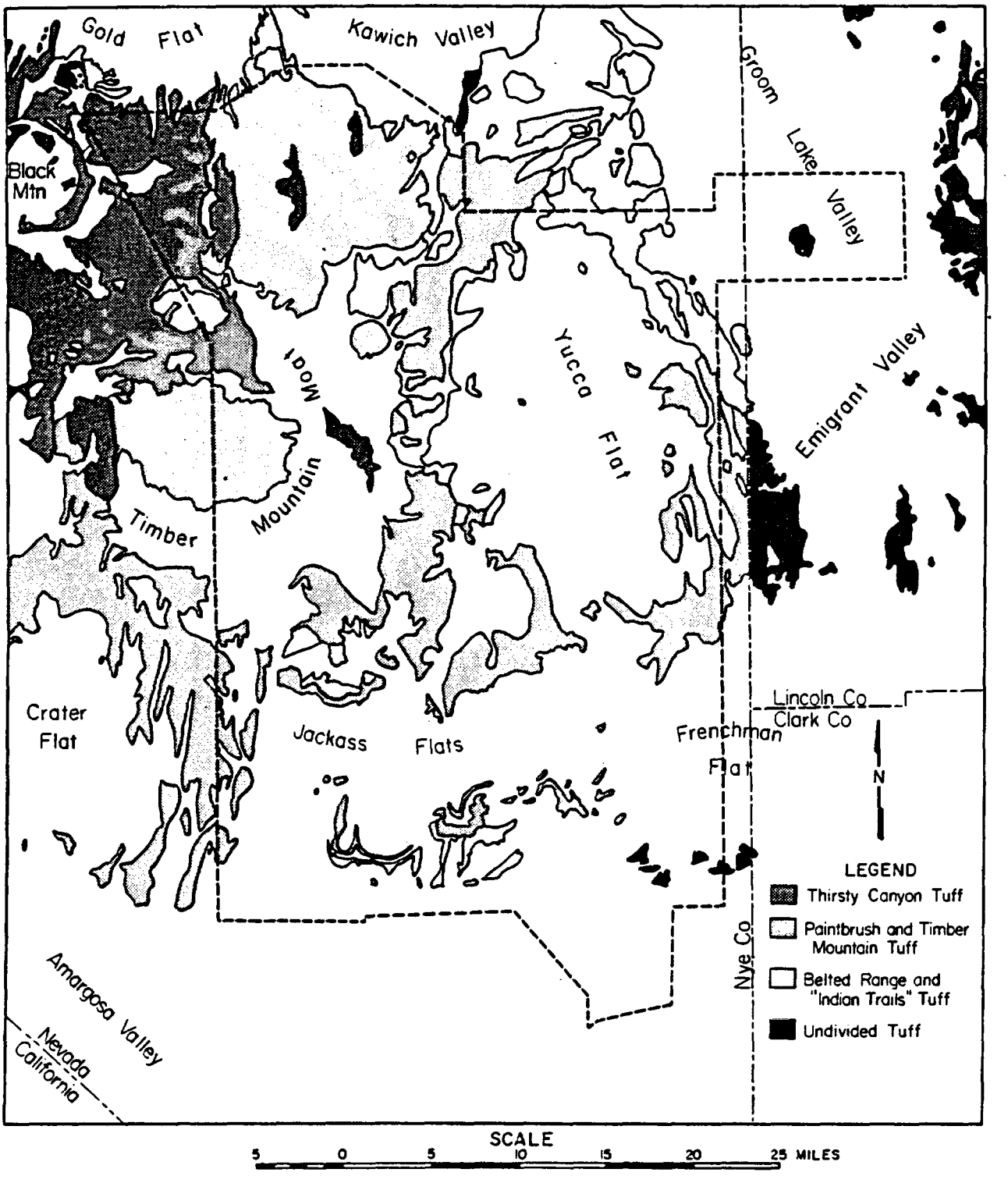


**Figure 17.** Distribution of Silicic Tuff and Related Rock Outcrops in the NTS Area (from Cornwall, 1972, Tschanz and Pampeyan, 1970, Longwell, et al., 1965b, and USGS GQ Map Series of the NTS Region)

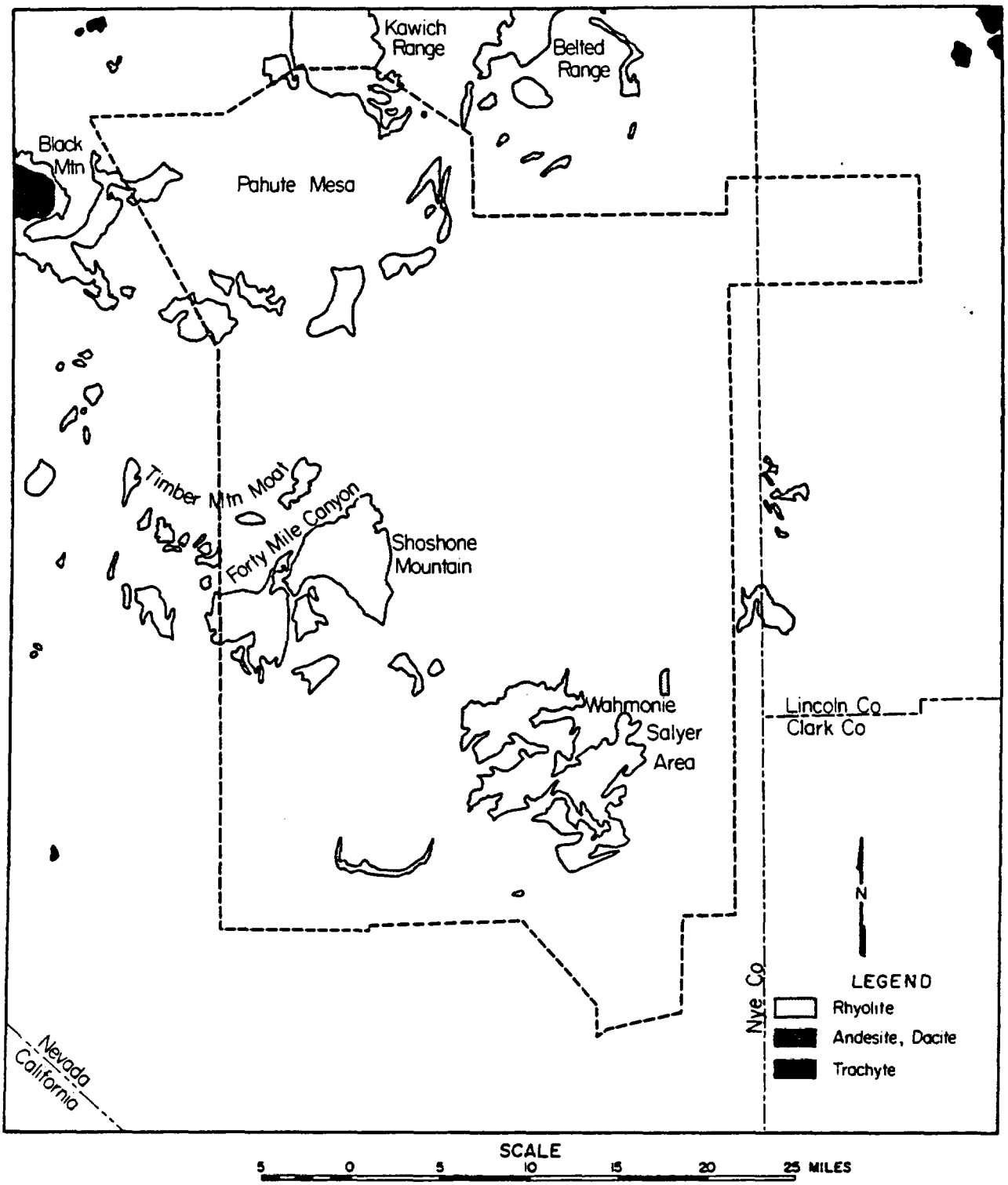


**Figure 18.** Distribution of Silicic and Intermediate Composition Lava Flows in the NTS Area (from Cornwall, 1972, Tschanz and Pampeyan, 1970, Longwell, et al., 1965b, and USGS GQ Map Series of the NTS Region)

## Monotony Tuff

The oldest major tuff deposits in the area form the Monotony Tuff, dated at 26-28 m.y. (Ekren et al., 1971). These deposits blanketed an area east and north of the NTS with the distal parts of some flows reaching as far west and south as the NTS. Other named units of the earliest series of southern Nevadan volcanic rocks are the Antelope Springs Tuff (dated as 26-27 m.y. old, by Ekren et al., 1971) and the White Blotch Spring Tuff (dated as 24-25 m.y. old, by Ekren et al., 1971; 22.3 m.y., old by Anderson and Ekren, 1968). Ekren et al., (1971) propose that the Monotony Tuff was extruded from a presently obscured caldera complex located at the southern end of the Pancake Range in central Nevada. The White Blotch Spring and Antelope Springs Tuffs probably originated in three or more distinct centers in the vicinity of the Cactus Range, the Kawich Range, and Mt. Helen (Figure 19). Associated deposits include minor rhyolites and sedimentary layers.

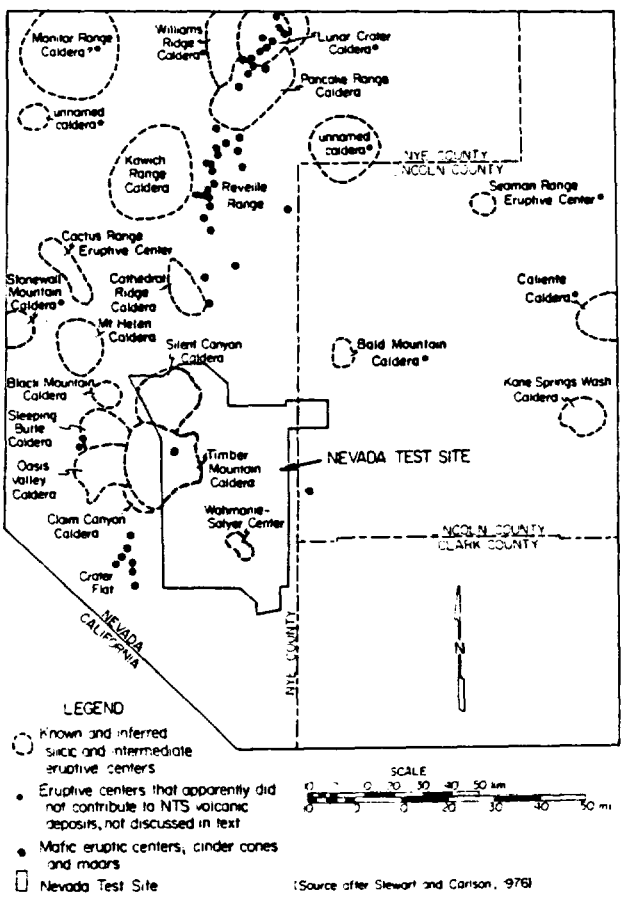


**Figure 19.** General Location of Known and Inferred Eruptive Centers in the NTS Vicinity (after Stewart, 1980)

The Monotony Tuff approaches 4000 ft in thickness in the Pancake Range. More significant is its total volume, estimated at more than 1000 cubic mi by Ekren (1968). The volume of White Blotch Spring Tuff is estimated as more than 500 cubic mi (Ekren, 1968) and approaches 3000 ft in maximum thickness in the Cactus Range (Ekren et al., 1971). The Antelope Springs Tuff is reported as 8600 ft thick in the Cactus Range (Ekren et al., 1971). No estimate of its total original volume is available.

Just north of the NTS a suite of intermediate composition igneous rocks overlies the oldest tuffs. This suite of rocks is composed of dactites, andesites, and quartz latites (Anderson and Ekren, 1968). These lavas, sills, and tuffs may account for as much as 3000 ft of composite thickness, but they are difficult to distinguish as discrete units because of complex interlayering and a paucity of diagnostic marker horizons (Ekren et al., 1971). The age of these rocks ranges from about 17.8 to 22.3 m.y. (Anderson and Ekren, 1968; Ekren et al., 1971). Source areas were numerous and widely distributed, though Mt. Helen apparently served as a recurrent center of extrusion (Ekren et al., 1971).

## Fraction Tuff

Overlying the intermediate composition lavas is the Fraction Tuff, the second major sequence of predominantly silicic ash-flow and air-fall tuff in the NTS area. This unit is dated as about 15-18 m.y., or middle Miocene (Ekren et al., 1971; Nolan, 1930). Ekren (1968) estimates its total volume as greater than 500 cubic mi. A composite thickness of 7200 ft is reported at Cathedral Ridge on the south end of the Kawich Range, the inferred venting source (Ekren et al., 1971). Thickness decreases outward from Cathedral Ridge to less than 1000 ft on Pahute Mesa and a few hundred feet in northern Yucca Flat. The Fraction Tuff rests on a surface of considerable paleotopographic relief (Ekren et al., 1971). This surface is presumed to have developed simultaneous with regional structural and topographic doming due to early Miocene igneous activity. Tuffaceous alluvium is interbedded locally between distinct cooling units of the Fraction Tuff.

After extrusion of the Fraction Tuff, continuing erosion produced local accumulations of conglomerate, siltstone, claystone, limestone, and tuffaceous sandstone. These sedimentary rocks overlie the Fraction Tuff in restricted areas north of the NTS (Ekren et al., 1971) and may indicate the beginning of Basin-Range normal faulting.

## Rhyolites and Tolicha Peak Tuff

Eruptions of several rhyolitic lavas, dated as 14-15 m.y. followed extrusions of the Fraction Tuff. These lavas include O'Brien's Knob, Cactus Peak, Belted Peak, and Ocher Ridge rhyolites (Ekren et al., 1971). Noble (1968) suggests that similar lavas of approximately the same age were precursors to voluminous ash-flow eruptions from the Kane Springs Wash volcanic center east of the NTS. Peralkaline (Na + K > Al) tuff was ejected from the Kane Springs Wash Caldera as a series of air-falls and ash-flows dated at about 14 m.y. (Noble, 1968). North of the NTS the Tolicha Peak Tuff was extruded at about the same time from the vicinity of Mt. Helen (Ekren et al., 1971). This unit is about 200-400 ft thick and covers broad areas adjacent to Mt. Helen. The dates and stratigraphic position of these units suggest their correlation with the Indian Trails Tuff Formation of the eastern NTS (Poole and McKeown, 1962).

The volcanic units discussed to this point were produced from volcanic centers located at some distance from the NTS. After the Tolicha Peak-Kane Springs Wash eruptions, eruptive centers shifted to areas on and near the NTS. The resulting deposits are generally less deformed and less hydrothermally altered than those erupted earlier. This allows better interpretation of eruptive histories, locations, and extents of the younger flow units, as well as better identification of caldera source areas for individual deposits. It is likely that the local history of each major tuff unit mentioned above was geographically and temporally as complex as the entire suite of younger tuffs described in the following paragraphs.

## Red Rock Valley and Crater Flat Tuffs

The Red Rock Valley Tuff is the oldest unit erupted from the vicinity of the NTS. It is reported as about 15-16 m.y. old by Marvin et al. (1970) and about 12-13 m.y. old by Kistler (1968). Byers et al., (1976a) estimate that its total volume is about 90 cubic mi, spread across 1200 sq mi with an average thickness of 400 ft. It may have erupted from the Sleeping Butte Caldera prior to the overlying and similarly calc-alkaline Crater Flat Tuff (Byers et al., 1976a), dated as about 14.0 m.y. (Marvin et al., 1970). Where they both occur, the Red Rock Valley and Crater Flat Tuffs are commonly separated by a few hundred feet of bedded, tuffaceous alluvium. A total volume of 250-300 cubic mi for Crater Flat Tuff, mostly the Bullfrog Member, is estimated by Byers et al. (1976a). Recent drilling at Yucca Mountain and Crater Flat has shown significant thickness of the Crater Flat Tuff, especially the lower Tram member, suggesting a source area in or near Crater Flat (Spengler et al., 1981).

The Crater Flat Tuff is locally overlain by the Tolicha Peak Tuff on the northern flank of Pahute Mesa. Thus, the general time of the Crater Flat-Tolicha Peak eruptions represents the transition of volcanic centers from the north to those on and west of the NTS.

## Belted Range Tuff

After deposition of the Crater Flat Tuff, peralkaline tuffs of the Belted Range Formation were extruded from the Silent Canyon Caldera at Pahute Mesa. This caldera is presently obscured by the younger Timber Mountain Tuff that forms the surface of Pahute Mesa (Orkild et al., 1968; Noble, 1970; Noble et al., 1968). The Belted Range Tuff has been dated as about 13-15 m.y. (Noble et al., 1968; Marvin et al., 1970). The volume of the lower Tub Spring Member is estimated as about 25 cubic mi, and has an average thickness of 300 ft spread across 1000 sq mi. The upper Grouse Canyon Member is more voluminous (50 cubic mi), much more widespread (3000 sq mi) and also averages about 300 ft thick (Noble et al., 1968). Byers et al., (1976a) distinguish the Stockade Wash Tuff of about 5-10 cubic mi as a separate formation and suggest that it is associated genetically with the Silent Canyon Caldera. Previous workers (Hinrichs and Orkild, 1961; Ekren et al., 1971; Poole and McKeown, 1962) considered the Stockade Wash Tuff as the basal member of the Paintbrush Tuff, the formation overlying the Belted Range Tuff.

## Paintbrush Tuff

The Paintbrush Tuff probably erupted from the Claim Canyon Caldera. It is composed of, from bottom to top, the Topopah Springs, Pah Canyon, Yucca Mountain, and Tiva Canyon Members (Byers et al., 1976a,b). More than 300 cubic mi of Paintbrush Tuff were extruded, with perhaps 200 cubic mi occurring outside the caldera basin (Byers et al., 1976a). The uppermost Tiva Canyon Member is the most widespread and voluminous, covering perhaps 1200 sq mi and containing up to 250 cubic mi of extrusive material (Byers et al., 1976a). A total thickness of 6500 ft of Paintbrush Tuff is reported within the Claim Canyon Caldera (Byers et al., 1976a). Composition of the Paintbrush Tuff varies from rhyolitic to calc-alkalic. Some individual cooling units exhibit zonal variations from bottom to top (Lipman and Christiansen, 1964;

Lipman et al., 1966a). The Paintbrush Tuff has been dated as about 12-13 m.y. (Kistler, 1968; Marvin et al., 1970). The upper two members of the Paintbrush Tuff may have originated from the Oasis Valley Caldera, distinct from but geographically overlapping the Claim Canyon Caldera (Christiansen et al., 1977).

Below the Paintbrush Tuff is a series of local lava flows that emerged along a system of fissures radial to the Claim Canyon Caldera. These fissures produced the Rhyolites of Calico Hills and related flows on Pahute Mesa (Christiansen et al., 1977). At Yucca Mountain, the Calico Hills units are composed of bedded and reworked air-fall tuff and nonwelded ash-flow tuff, about 300 ft-1000 ft thick (Spengler et al., 1981). Rhyolites of Calico Hills have been dated as 13.4 m.y. old (Kistler, 1968).

## Salyer and Wahmonie Rhyolites

Interbedded rhyolitic to andesitic lavas, breccia flows, and minor tuffs were also extruded from the Salyer and Wahmonie noncaldera, dome volcanoes at about the same time (12-13 m.y. ago as reported by Kistler, 1968, for the Wahmonie Formation). The Salyer Formation rests on Mara Wash Tuffs and the overlying Wahmonie Formation occurs below and interbedded with the Paintbrush Tuff (Poole et al., 1965a). The calc-alkalic Salyer and Wahmonie lavas (Christiansen et al., 1977) are believed to have originated from separate volcanoes, perhaps with a common magma chamber at depth. The larger Wahmonie flows covered about 500 sq mi and reached a maximum thickness of 3500 ft near Wahmonie Flat. The Salyer Formation covers about 300 sq mi and approaches 2000 ft in thickness at Mount Salyer.

## Timber Mountain Tuff

The last, and largest, of the great caldera eruptions centered near Timber Mountain exploded into activity after the Paintbrush eruptions. The lower Rainier Mesa and upper Ammonia Tanks members of the silicic Timber Mountain Tuff comprise about 500-550 cubic mi. Associated tuffs of Buttonhook Wash and Crooked Canyon add little additional volume (Byers et al., 1976a; Ekren, 1968). Within the Timber Mountain Caldera the total thickness of Timber Mountain Tuff and related rocks may exceed 5000 ft (Byers et al., 1976a,b). The area covered by Timber Mountain flows approaches 5000 sq mi. The age of the eruptions is about 9.5-11.5 m.y. (Kistler, 1968; Marvin et al., 1970; Christiansen et al., 1977). Rhyolites of Fortymile Canyon (Byers et al., 1976a,b; Christiansen et al., 1977) and mafic lavas of Dome Mountain (Luft, 1964) Kiwi Mesa, and Skull Mountain were extruded at the end of the Timber Mountain volcanic cycle and may be as young as 9 m.y. (Byers et al., 1976a).

## Thirsty Canyon Tuff

Following activity at Timber Mountain, one final pulse of silicic extrusions, the Thirsty Canyon Tuff, spread across the northwestern NTS about 6-8 m.y. ago from the Black Mountain Caldera (Kistler, 1968; Marvin et al., 1970; Crowe and Sargent, 1979; Noble et al., 1964). These tuffs surrounded the northwest sides and partially spilled into the basin of the Timber Mountain Caldera. However, the general topographic expression of the Timber Mountain Caldera is preserved today, as is the Black Mountain Caldera (Figure 20, Sections A-A' and C-C'). The predominantly peralkaline Thirsty Canyon Tuff is divided into five members, from oldest to youngest: Rocket Wash, Spearhead, Trail Ridge, Gold Flat, and Labryinth Canyon (Ekren et al., 1971). The total volume of all members is estimated at 50 cubic mi, spread across about 1500 sq mi (Ekren, 1968).

## Summary of Volcanism in NTS Area

The Thirsty Canyon Tuff represents the culminating phase of about 8-10 m. yr. of silicic eruptions in the NTS area. At least eight major mid-Miocene to Pliocene calderas supplied eruptive material to the NTS (Figure 19, Table 1). Five of these, the Sleeping Butte, Silent Canyon, Claim Canyon, Oasis Valley, and Timber Mountain, overlap geographically in the Timber Mountain-Pahute Mesa vicinity. This complex of eruptive centers may be considered a single, explosive volcano characterized by episodic eruptions originating perhaps from discrete parent magmas. The Cathedral Ridge, Kane Springs Wash, and Black Canyon calderas are outside the NTS boundaries but contributed volcanic materials to the area. Oligocene and early Miocene silicic eruptions occurred north and west of the NTS at Mt. Helen, the Cactus Range and the Northern Kawich Range. Some flows from these earlier volcanoes reached the northern edge of the NTS. The greatest thickness of Tertiary volcanic rocks is confined to intracaldera regions where it may exceed 15,000 ft, particularly in the vicinity of the Timber Mountain (Byers et al., 1976a,b) and Silent Canyon calderas (Orkild et al., 1968) (Figure 20, Section C-C'). Immediately outside the calderas, the thickness of the Tertiary volcanic pile locally approaches 10,000 ft, but it generally thins to extinction toward the south and southeast portions of the NTS (Figure 20, Section A-A').

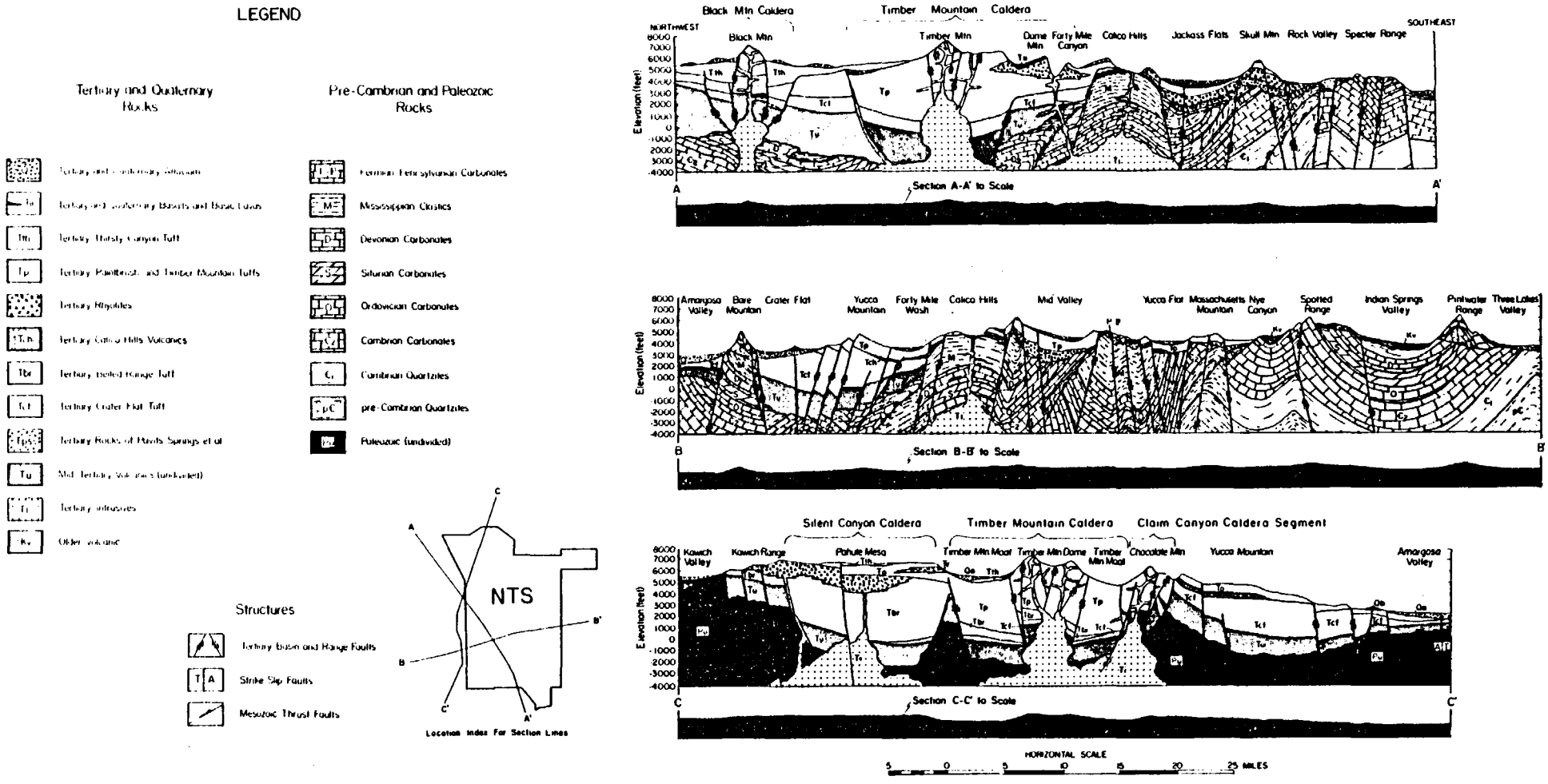


**Figure 20.** Idealized Structure Sections of the NTS Area (Vertical Exaggeration is by a Factor of 5); Darkened Topographic Sections Below Each Structure Section are to Scale

## Table 1. Volcanic Stratigraphy of the NTS Area

| EPOCH | ERUPTIVE CENTER | FORMATION | MEMBERS OR RELATED UNITS | RADIOMETRIC AGE-$10^6$YRS | ESTIMATED VOLUME-$MI^3$ |
|---|---|---|---|---|---|
| PLIOCENE | BLACK MOUNTAIN CALDERA | THIRSTY CANYON TUFF | LABYRINTH CANYON MEMBER<br>GOLD FLAT MEMBER<br>TRAIL RIDGE MEMBER<br>SPEARHEAD MEMBER<br>ROCKET WASH MEMBER | 6-8 | 50 |
| | TIMBER MOUNTAIN CALDERA | TIMBER MOUNTAIN TUFF | RHYOLITES OF SHOSHONE MOUNTAIN<br>MAFIC LAVAS OF DOME MOUNTAIN<br>RHYOLITES OF FORTY MILE CANYON<br>TUFFS OF CROOKED CANYON AND BUTTONHOOK WASH<br>AMMONIA TANKS MEMBER<br>RAINIER MESA MEMBER | 9.5-11.5 | ? BUT SMALL<br>? BUT SMALL<br>? BUT SMALL<br>230<br>300 |
| MIOCENE | WAHMONIE - MT. SALYER AREA | WAHMONIE FORMATION<br>SALYER FORMATION | MULTIPLE RHYOLITE, ANDESITE AND BRECCIA FLOWS, AND THIN TUFFS | 12-13 | 50? |
| | CLAIM CANYON CALDERA - CALICO HILLS | PAINTBRUSH TUFF | RHYOLITE FLOWS<br>TIVA CANYON MEMBER<br>YUCCA MOUNTAIN MEMBER<br>LAVA FLOWS<br>PAH CANYON MEMBER<br>TOPOPAH SPRINGS MEMBER | 12-13 | ?<br>250<br>4<br>?<br>5<br>60 |
| | | CALICO HILLS FORMATION | RHYOLITES OF CALICO HILLS<br>ASHFLOW AND ASHFALL TUFFS | 13-14 | ? |
| | SILENT CANYON CALDERA | STOCKADE WASH TUFF | | 13-15 | 5-10 |
| | | BELTED RANGE TUFF | GROUSE CANYON MEMBER<br>TUB SPRINGS MEMBER | 13-15 | 75 |
| | CRATER FLAT AREA-SLEEPING BUTTE CALDERA | CRATER FLAT TUFF | PROW PASS MEMBER<br>BULLFROG MEMBER<br>TRAM MEMBER | 14-15 | 300 |
| | | RED ROCK VALLEY TUFF | | 14-16 | ? BUT SMALL |
| | MT. HELEN | TOLICHA PEAK TUFF | MULTIPLE COOLING UNITS | >14 | ? |
| | KANE SPRINGS WASH CALDERA | KANE WASH TUFF | MULTIPLE COOLING UNITS<br>LAVA FLOWS | 14-15 | 200 |
| | CACTUS-KAWICH RANGES | RHYOLITE LAVA FLOWS | O'BRIEN'S KNOB, CACTUS PEAK<br>BELTED PEAK, OCHER RIDGE | 14-15 | 200 |
| | CATHEDRAL RIDGE CALDERA | FRACTION TUFF | MULTIPLE COOLING UNITS | 15-18 | 500 |
| | MT. HELEN, CACTUS-KAWICH RANGES | FLOWS OF INTERMEDIATE COMPOSITION | DACITES, ANDESITES, QUARTZ LATITES | 18-22 | ? |
| | | WHITE BLOTCH SPRING TUFF | MULTIPLE COOLING UNITS | 24-25 | 500 |
| OLI-GOCENE | | ANTELOPE VALLEY TUFF | MULTIPLE COOLING UNITS | 26-27 | ? |
| | PANCAKE RANGE | MONOTONY TUFF | MULTIPLE COOLING UNITS | 26-28 | 1000 |

The age assigned to the inception of volcanism and individual volcanic units depends on: (1) the interpretations of individual investigators; (2) the reliability of correlations for a single formation throughout the region; (3) local differences in the time of inception of volcanism; and (4) the reliability of radiometric dating techniques. It is therefore difficult to unequivocally establish the location of many eruptive centers, cycles of multiple eruptions from individual centers, and geographical and temporal overlap of penecontemporaneous eruptions from separate areas. Notwithstanding, a general sequence of events for Tertiary volcanism has been established for the NTS region. The series of volcanic events outlined above led to the formation of about 75-100 discrete geological mapping units. The time span for this volcanism was about 15-20 m. y. (Figure 21). Many of the volcanic events are assigned a single age, though one or several million years may have been required to produce the resulting deposits.

## Cenozoic Intrusive Rocks

Intrusion of granitic stocks accompanied Tertiary volcanism in the NTS area (Spengler et al., 1979b). Small stocks are exposed east of Lookout Peak (the Wahmonie Stock) (Ekren and Sargent, 1965), and along the southeast and north edges of the Timber Mountain resurgent dome (Byers et al., 1976b). A granite was encountered at a depth of about 9000 ft in drill hole PM2. This granite intrudes volcanic rocks at the northwestern corner of the NTS (Orkild et al., 1969). Other Tertiary granitic stocks are inferred to occur below Calico Hills (Christiansen et al., 1977, Snyder and Oliver, 1981) and the Timber Mountain Caldera (Byers et al., 1976a; Kane et al., 1981). Numerous rhyolitic and basaltic dikes, plugs, and sills are exposed throughout the NTS. The distribution of intrusive Tertiary rocks in the NTS area is shown in Figure 16 with the Mesozoic plutons.

## Late Pliocene and Pleistocene Basalts

After eruption of the Thirsty Canyon Tuff 6-8 m.y. ago, silicic volcanism was succeeded by small, predominantly basaltic eruptions. Although not synchronous, a general transition from large silicic to small basaltic eruptions occurred throughout the Great Basin. With exception of the Bishop Tuff from Mono County, California, dated as 0.7 m.y. (Dalrymple et al., 1965), no Pleistocene silicic tuffs are known by the author to occur in the Great Basin of Nevada. However, Upper Pliocene and Pleistocene basalts and related rocks are common. The transition from predominantly tuff to basalt eruptions occurred later in the NTS area, about 6-7 m.y. ago, than along the western and northern margins of the Great Basin where a similar transition occurred about 12-16 m.y. ago (Snyder et al., 1976; McKee, 1971; Noble, 1972).

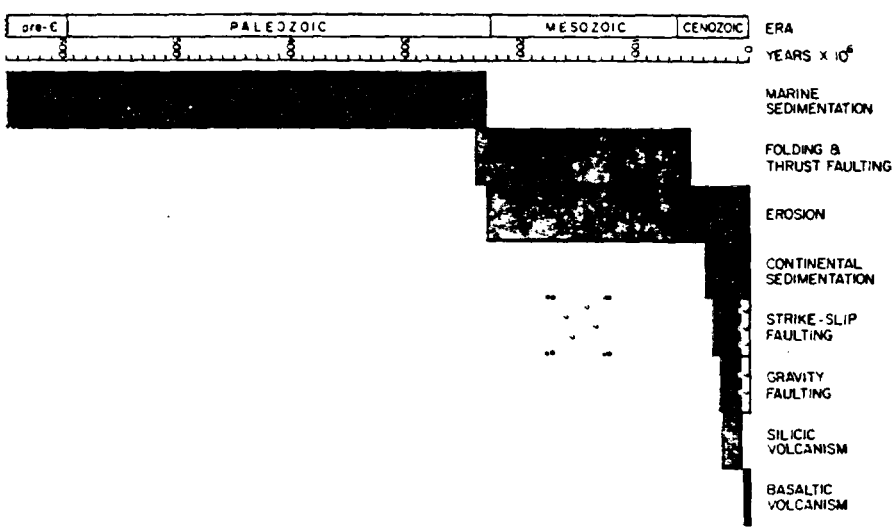


**Figure 21.** Time Distribution of Major Geologic Events in the NTS Region

Upper Tertiary and Quaternary volcanic rocks in the NTS region are predominantly alkali-olivine basalts or hawaiites, with minor basaltic-andesite differentiates (Leeman and Rogers, 1970; Hausel and Nash, 1977; Best and Brimhall, 1974; Vaniman et al., 1980). Though extrusion of basalts accompanied Miocene and early Pliocene silicic eruptions in the NTS area, the earlier basalts were late stage, silica saturated trachy basalt and andesite differentiates of the silicic parent magma (Luft, 1964; Byers et al., 1976a; Armstrong, 1970). The younger basalts are probably derived from deeper, subcrustal magma sources and suggest intense lithospheric extension throughout the Basin and Range (McKee, 1971; Best and Brimhall, 1974; Snyder et al., 1976; Crowe et al., 1980; Vaniman et al., 1980).

In the NTS area young alkalic-olivine basalts occur in Crater Flat (Crowe and Carr, 1980; Cornwall and Kleinhampl, 1961), the Timber Mountain Moat (Byers et al., 1976b), Kawich Valley (Ekren et al., 1971), the southern end of the Halfpint Range (Byers and Barnes, 1967), and along Rocket Wash (Lipman et al., 1966b) (Figure 22). These basalts range from about 5 m.y. old to as young as perhaps 0.25 m.y. (Sinnock and Easterling, 1982). In Crater Flat ages have been determined of about 1-1.5 m.y. for Black Cone, Red Cone (Figure 23), and, by inference, Little Cone; .1-.75 m.y. for Lathrop Wells Cone (Figure 24) (Carr, oral communication; Sinnock and Easterling, 1982); and about 4.0 m.y. for a series of flows in the eastern portion of the valley (Sinnock and Easterling, 1982). Basalt from Buckboard Mesa in the Timber Mountain moat has been dated as about 2.5-3.0 m.y. old (Carr, oral communication).

Other Plio-Pleistocene basalts of the NTS area, except perhaps some basalts in Kawich Valley and Sarcobatus Flat, are believed to be older than those already dated. Basalt flows in Crater Flat, Kawich Valley and dikes and sills along Scarp Canyon and Nye Canyon of the Halfpint Range and upper Frenchman Flat (Figure 25) rest on or intrude alluvial deposits. Some exhibit evidence of cinder cones, thereby indicating relatively young ages. Carr (1974) reports that basalt sills and dikes intrude fault zones and alluvial deposits of Yucca Flat. Drilling has encountered basalt in alluvium below the surface of Crater Flat. Study by the U.S. Geological Survey and Los Alamos National Laboratory of the age and distribution of Pleistocene basalt in the NTS area is currently underway.

## Late Pliocene Through Recent Alluvium

Though normal faulting in the NTS area probably began during early phases of silicic volcanism, current Basin and Range topography did not begin to take form until later phases of silicic eruptions (Ekren et al., 1968). At this time, topographically closed basins were formed, and material eroded or extruded from the ranges began accumulating in the basins. Alluvial deposition has continued essentially uninterrupted to the present, whereas episodic volcanic basin filling essentially ceased about 6-8 m.y. ago. Hinrichs (1968b), Fernald et al. (1968b) and Grothaus and Howard (1977) report about 2000 ft of alluvium in parts of Yucca Flat. McKay and Williams (1964) estimate about 1000 ft of alluvium in Jackass Flats near the EMAD facilities and infer that it entirely postdates Kiwi Mesa Basalt.

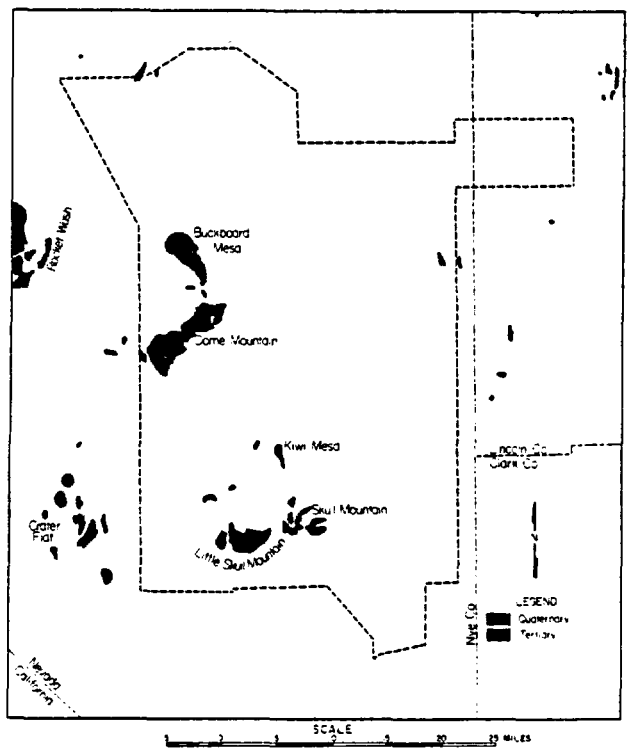


**Figure 22.** Distribution of Late Tertiary and Quaternary Basalt Flows in the NTS Area (from Cornwall, 1972, Tschanz and Pampeyan, 1970, Stewart and Carlson, 1976, and USGS GQ Map Series of the NTS Region)

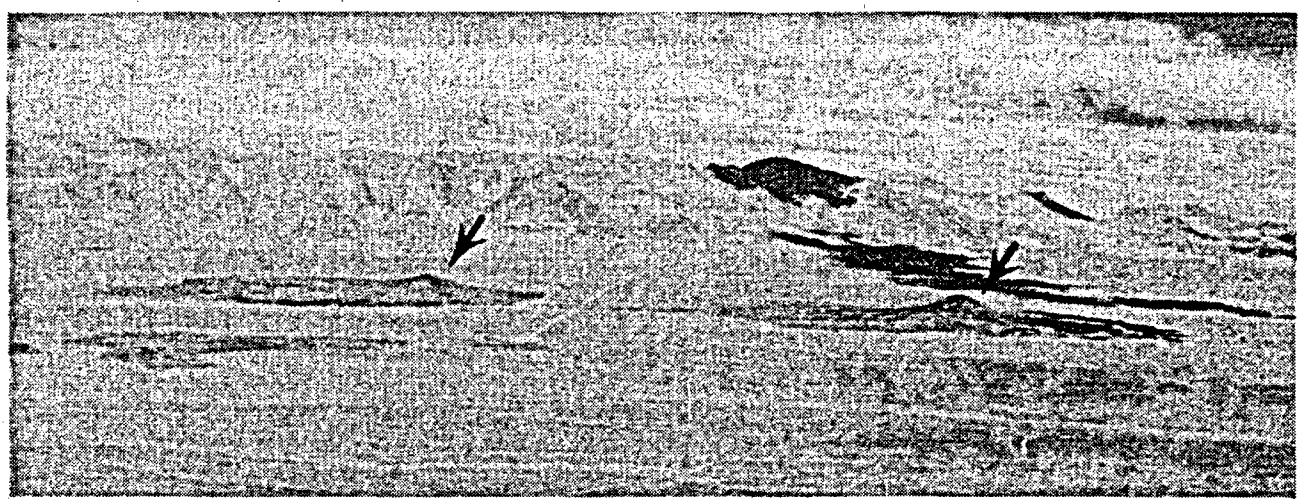

**Figure 23.** Black Cone (Right Arrow) and Red Cone (Left Arrow) and Associated Basalt Flows in Crater Flat, Nevada. Bare Mountains in background. Oblique aerial view is to the west from the southwest edge of the NTS. Photo courtesy of Pan American, Inc.

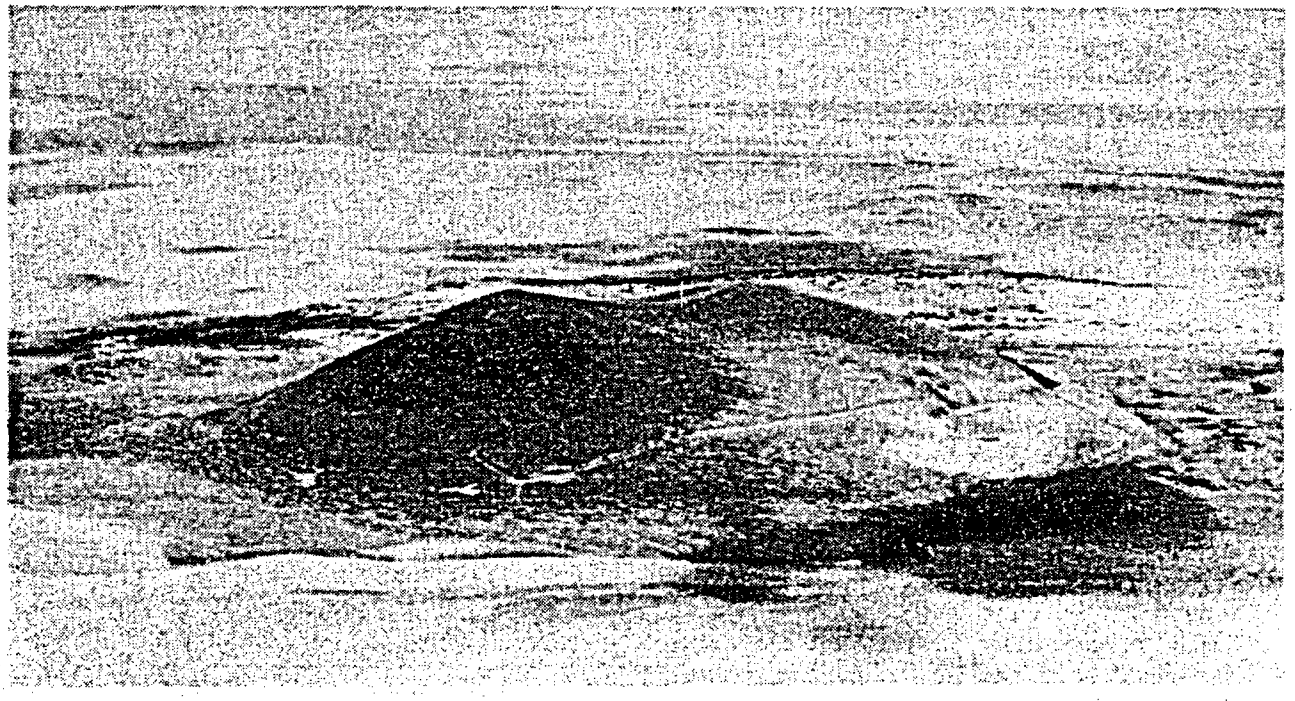

**Figure 24.** Volcanic Cinder Cone Along U.S. 95 Seven Miles West of Lathrop Wells, Nevada. Note unbreached crater. Age of the cone and associated basalt flows is less than 0.75 million years. Oblique aerial view is looking northeast. Photo courtesy of Pan American, Inc.

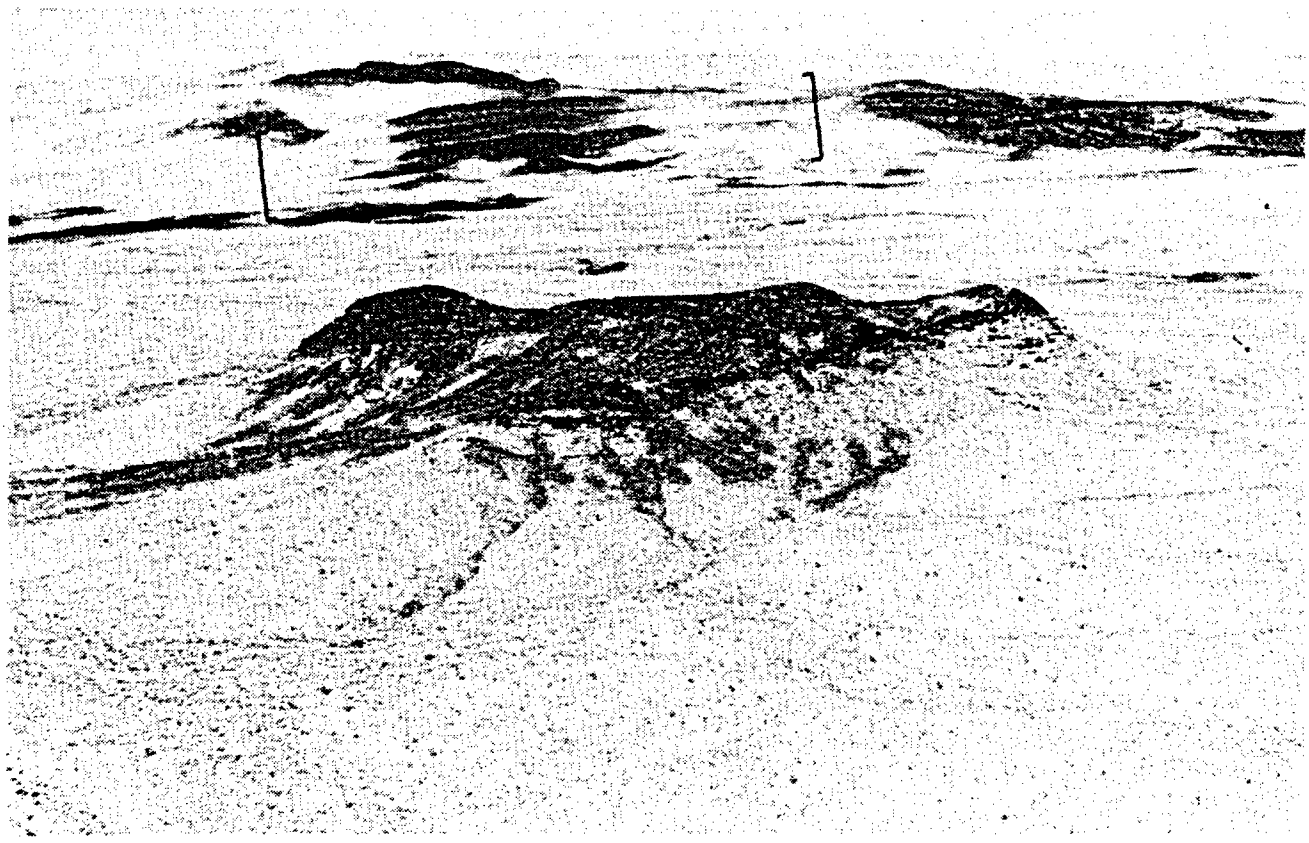

**Figure 25.** Basalt Dikes and Sills in the Upper Reaches of Frenchman Flat Just East of the NTS. Note parallel, north trending fault blocks in the Halfpint Range (brackets). Oblique aerial view is looking northwest. Photo courtesy of Pan American, Inc.

The alluvium is generally unconsolidated. Some layers are partially to firmly indurated by calcium-carbonate (caliche) and/or clays. The alluvial materials are better sorted and finer grained toward basin centers (Fernald et al., 1968a, 1968b). The playas consist of very fine-grained lacustrine deposits up to several hundred feet thick. Near the range fronts alluvium is generally composed of angular rubble, with individual clasts commonly a foot or more in diameter surrounded by a matrix of silt, sand, and gravel. Figure 26 shows the distribution of alluvial deposits in the NTS area.

Scant precipitation and intense evaporation in the arid Great Basin prohibit development of large perennial streams capable of transporting eroded sediment away from local source areas. Alluvial fans tend to occur along range fronts where sediment is dropped as stream gradients abruptly flatten. Fans from opposing ranges commonly coalesce at the lower parts of the basins. Gradients and surface areas of individual fans depend on the available relief in the ranges, lithologic composition of the ranges, drainage area of the headwaters, and abundance and periodicity of precipitation.

# Rock Structures

## General Statement

Rock structures on the NTS may be divided into regional, volcanic, and local structures, including man-made features. Regional structures are caused by widely distributed tectonic body stresses and include (1) compressional features, (2) extensional features, (3) northwest-southeast strike-slip features, and (4) northeast-southwest strike-slip features. Volcanic structures are caused by local vertical stresses from upward movement of heat and magma and include concentric and radial fracture systems associated with volcanic doming, caldera collapse, and caldera resurgence. Local structures are created by the responses of local rock inhomogeneities to regional tectonic stresses, volcanic and intrusive processes, sediment compaction, sediment dessication, erosional unloading, weathering, underground explosions, hydraulic fracturing, mining, drilling, and others. Minor structures on the NTS include small faults, joints, fractures, underground cavities, and open cracks.

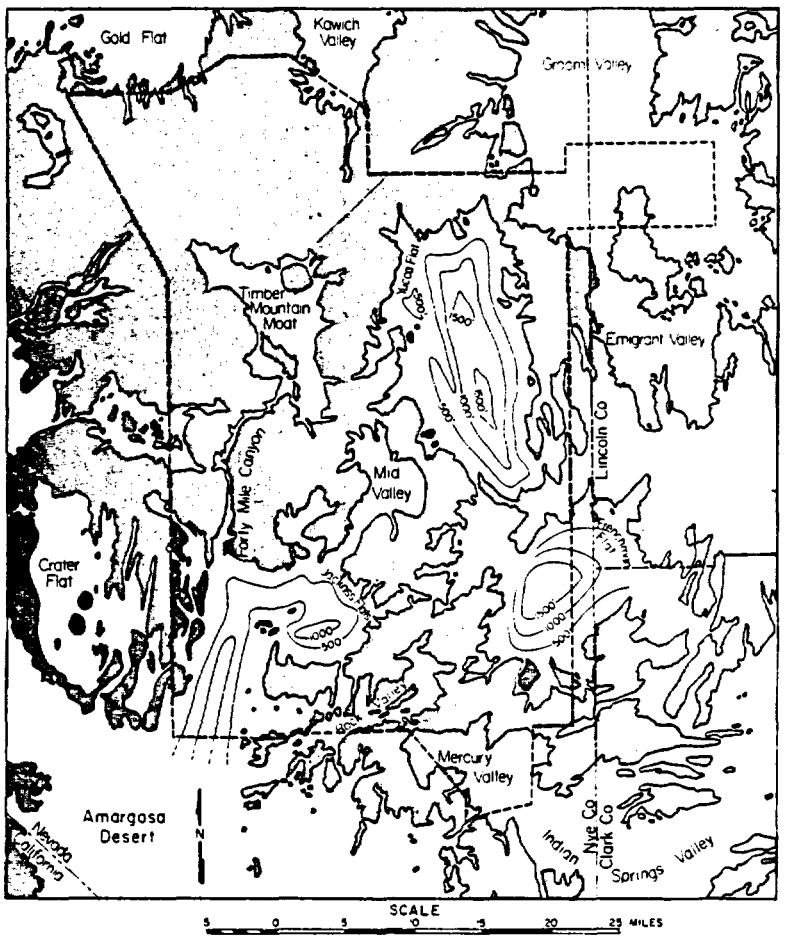


**Figure 26.** Distribution of Alluvium (Unshaded) in the NTS Area; Contours Show Inferred Thickness of Alluvium in Feet for Portions of Frenchman, Yucca, and Jackass Flats (from Cornwall, 1972, Tschanz and Pampeyan, 1970, Longwell, et al., 1965b, and USGS GQ and I Map Series for the NTS Region)

Whether rocks subjected to high stresses will fracture (fault) or deform in a ductile fashion (fold) depends on many factors, not the least of which are time, temperature, confining pressure, and the mechanical properties of the deformed media. If a given force is applied rapidly to a material, fracture tends to occur, whereas application of the same or even a reduced force over a longer time period may result in folding. Similarly, higher temperatures and confining pressures tend to cause folding rather than faulting under similar stresses. Therefore, faults tend to form more readily near the surface where temperatures and pressures are relatively low, whereas folds tend to be the dominant mode of deformation deeper within the crust where ductile behavior is induced by higher temperatures and pressures.

Given the complex relations among factors that affect the type and geometry of geological deformation, it is difficult to specify the exact nature of a local stress environment during a deformational period (orogeny) or to determine the exact sequence of events in a region characterized by many styles and/or periods of rock deformation.

## Mesozoic Compressional Structures

Compressional structures are produced if body shear stresses exceed the elastic strain limit of rocks. Considering the variety of rock types present at the NTS, it is apparent that a regional compressive force would create a variety of structural features, depending on the mechanical properties and geometric distribution of the various rock types. Stratigraphic evidence suggests that the NTS region was affected by more than one compressional mountain building episode during latest Paleozoic through earliest Cenozoic time, though correlations of individual structures with particular episodes are subject to alternate viewpoints.

At least one and perhaps two Paleozoic orogenies, the Antler and Sonoman, affected the area northwest of the NTS. These mountain building episodes created a land mass and an eastward bounding sedimentation trough where the Eleana, Tippipah, and Bird Springs formations were deposited (Churkin, 1974; Poole, 1974; Bissell, 1974; Burchfiel and Davis, 1975). These upper Paleozoic rocks are deformed by compressive structures; the younger volcanic rocks are not. Thus, after deposition of the Bird Springs and Tippipah limestones in Permian time and before extrusion of the earliest volcanic rocks in Tertiary time, the NTS area experienced intense compressional stresses. These stresses were generally directed west to east and northwest to southeast. The resulting structures are two types; folds, commonly recumbent, and faults, predominantly thrusts.

## Thrust Faults

Compressive mountain building generally migrated to the east and southeast in Nevada during the Mesozoic Era (Burchfiel and Davis, 1975; Dott and Batten, 1971; Fleck, 1970; Armstrong, 1968b; Bissell, 1974). Burchfiel et al. (1970), place the time of initial thrust faulting and folding in the NTS area during the middle Triassic to Early Jurassic Nevadan Orogeny. Three major and numerous minor thrust faults formed in the NTS area at this time. From west to east, major thrust faults are designated as the CP, Mine Mountain, and Tippinip Faults by Barnes and Poole (1968), Ekren et al. (1971), and Ekren (1968). The CP structure is interpreted as a high angle "parent" thrust plane beneath the Belted Range. The Mine Mountain and Tippinip thrusts are thought to represent low angle, near surface imbrication in front of the CP thrust root zone (Barnes and Poole, 1968). Total lateral displacement along the fault system is estimated as 25-35 mi (Barnes and Poole, 1968). If indeed the CP structure has a deep origin, this implies that the underlying crust has shortened, presumably by ductile behavior.

The CP structure lifts and translates Cambrian through Devonian rocks over Mississippian and Pennsylvanian rocks. The base of the upper thrust sheet is formed by Devils Gate Limestone in Calico Hills and Eureka quartzite near Mine Mountain. The readily deformed, ductile Eleana Formation provided the basal slip surface.

The Mine Mountain thrust sheet places Cambrian rocks over Pennsylvanian and Mississippian rocks near the southwestern edge of Yucca Flat and pre-Cambrian and Cambrian clastic rocks above Cambrian carbonates southwest of Groom Lake. However, in most places the Mine Mountain thrust generally exhibits little stratigraphic displacement and commonly is limited to movement entirely in Devonian-Mississippian rocks (Barnes and Poole, 1968).

The Tippinip thrust places Cambrian over Ordovician carbonates in the Specter Range and Spotted Range. Carbonates in the lower plate are recumbently folded (Barnes and Poole, 1968). Burchfiel (1965) suggests thrusting in the Specter Range area is related to decollement along the Johnnie Thrust of the Spring Mountains (Nolan, 1929). Longwell (1974) and Armstrong (1968b) assign the Johnnie Thrust to the Late Mesozoic Sevier Orogeny.

The pattern of exposed thrust faults in the NTS area is shown on Figure 27. Because thrust fault surfaces are commonly imbricate and nearly horizontal, erosional incision in thrusted terrain may expose an exceedingly complex pattern of surface traces along the fault system. This makes it difficult to assign individual fault exposures to discrete regional thrust planes. Interpretation is required to define the simplest geometry that satisfies the evidence. An alternative interpretation of the thrust zones is proposed in this report wherein each of the major thrust zones possesses its own root, deeply seated within the upper or middle crust (Figure 20, Sections A-A' and B-B'). Previous interpretations have assumed one deep-seated root zone for all thrusting in the NTS area. However, outcrop patterns of the thrust faults at the NTS seem to lie in three distinct belts, suggesting separate roots (Figure 27). The geographic locations of the three belts suggest their names should be modified to, from west to east: the Mine Mountain, CP-Tippinip, and Spotted Range thrust zones (Figure 27). Though evidence is scant regarding either the single or multiple root-zone interpretations, the geographic pattern of thrust fault outcrops shown in Figure 27 suggests multiple root zones, perhaps merging somewhere in the middle to upper crust.

## Folds

The complexity of Mesozoic deformation is accentuated by folds formed by the same compressive stresses responsible for the thrust faults. Fold axes in the region generally trend north to northeast. The only named fold in the area is the Carbonate Wash Syncline just north of the NTS boundary in the Belted Range (Rogers and Noble, 1969). A large anticline is developed in Eleana quartzites at Quartzite Ridge (Barnes et al., 1963). A large syncline occurs at Syncline Ridge along the western edge of Yucca Flat. Carr (1974) maps several synclines and anticlines in the Yucca Flat area, including a large anticline at the

northern end. At other locations where Paleozoic rocks are exposed, Cenozoic and Mesozoic faulting has obscured the fold axes in Paleozoic rocks, though stratigraphic dips of 20° and greater suggest a ubiquitous folded character of the miogeosynclinal strata.

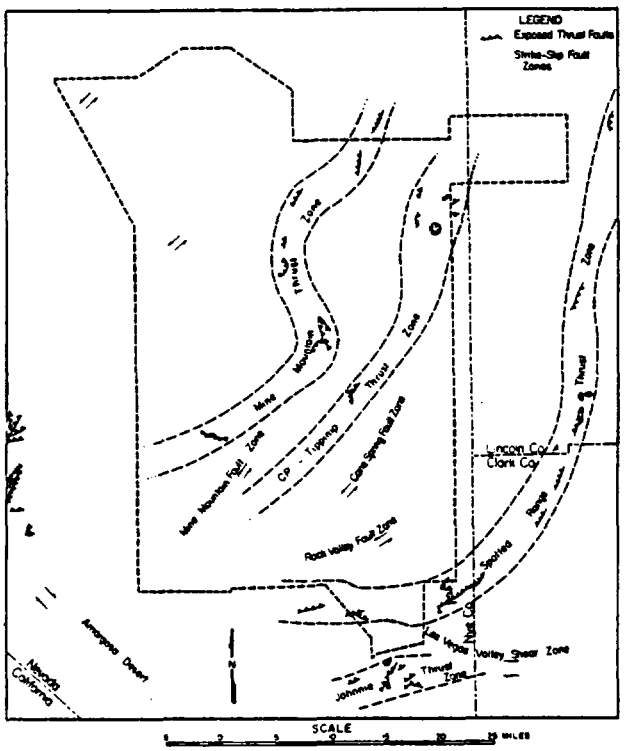


**Figure 27.** Outcrop Distribution of Thrust Fault and Strike-Slip Zones in the NTS Area (from Cornwall, 1972, Tschanz and Pampeyan, 1970, Longwell, et al., 1965b, and USGS GQ Map Series of the NTS Region)

## Sevier Orogeny

The Sevier Orogeny caused similar folding and thrusting during late Mesozoic time in areas east of the NTS (Harris, 1959; Armstrong, 1968b; Burchfiel and Davis, 1972, 1975). The Middle to Late Cretaceous-Early Tertiary age of the Sevier mountain building episode is better established than the ages of earlier orogenies. This is because clastic deposits derived from the ancient Sevier highland are well preserved in the Indianola Group and related rocks of Utah and Central Nevada (Spieker, 1949; Stokes, 1960, Cobban and Reeside, 1952) and in the Baseline Sandstone, Willow Tank Sandstone, Overton Fanglomerate and related rocks of southern Nevada (Longwell, 1952; Longwell et al., 1965a; Armstrong, 1968b). These rocks were spread eastward from the Sevier highlands in a manner analogous to the formation of Eleana deposits from erosion of the late Paleozoic Antler Highlands.

Longwell (1960) mapped six distinct thrust faults in the Spring Mountains associated with the Sevier Orogeny. At least two, the Wheeler Pass and Keystone thrusts, are exposed north of the Las Vegas Valley in the Muddy Mountains, where they are named the Gass Peak and Glendale-Muddy Mountain thrusts, respectively (Longwell et al., 1965a; Armstrong, 1968b; Stewart and Poole, 1974; Fleck, 1970; Brock and Engelder, 1977). Burchfiel and Davis (1977) trace the Keystone thrust south to the Clark and New York Mountains of California, and Drewes (1978) traces the Sevier compressional style from there into Mexico. To the north, Sevier deformation extends through Utah, Wyoming, Montana (Armstrong, 1968b) and British Columbia (Eisbacher et al., 1974).

No new structures are thought to have formed at the NTS during the Sevier Orogeny, though it is likely that movement along existing thrust planes and folds was rejuvenated. It appears certain that a wave of compressive deformation migrated eastward in the southern Nevada area during late Paleozoic and Mesozoic time. The Sevier pulse of this deformational wave culminated east of the NTS during latest Cretaceous time. Mountain building continued in the western United States after Sevier time (Compton et al., 1977), but the focus shifted to the Rocky Mountain area. The latter deformation is generally referred to as the Laramide Orogeny (Armstrong, 1968b; Anderson and Picard, 1974) and occurred from very latest Cretaceous through Eocene time.

# Cenozoic Volcanic Structures

Following compression and subsequent erosion of the Great Basin area during the Mesozoic and early Tertiary, volcanic material was spread across the region in middle and late Tertiary time as described earlier. Structures associated with the volcanic centers merit distinction. Each caldera tended to develop an initial broad dome with radial fractures, sets of concentric normal faults due to collapse of the central part of the dome, a circular collapse moat, and in some cases, an inner resurgent dome with a set of radial extension faults (Smith and Bailey, 1968). In the NTS area, Timber Mountain and Black Mountain calderas prominently display these features (Byers et al., 1976a; Carr, 1964; Christiansen et al., 1965; Carr and Quinlivan, 1968) (Figure 20, Sections A-A' and C-C').

At Timber Mountain most collapse features of the outer moat and resurgent ring-fractures of the inner moat are buried by late phase flows (Christiansen et al., 1977). Displacements along the outer edge of the caldera due to moat collapse may exceed 5000 ft (Christiansen et al., 1977). These outer ring fractures formed as the caldera collapsed into a subsurface

cavity created during extrusion of the volcanic material (Smith and Bailey, 1968). Radial faults and concentric inner-moat ring-fractures formed subsequently by near surface extension that accompanied resurgence of a central dome. Displacement along the concentric, inner fracture zone at Timber Mountain is about 1000 ft (Byers et al., 1976a,b). Radial faults approach 2000 ft of offset near the center of the dome but decrease outward (Carr and Quinlivan, 1968). A large complex of grabens along the crest of Timber Mountain generally trends east-west and is flanked by minor horsts and grabens typical of resurgent domes (Carr and Quinlivan, 1968). The inner ring-fractures and radial faults were loci for late phase intrusion and minor extrusion.

Timber Mountain and Black Canyon calderas are the only volcanic centers in the NTS area that display a wide range of typical structures. It is assumed, however, that such structures are present at most caldera centers defined earlier, but are now obscured by younger volcanic deposits and faulting. The Wahmonie, Salyer, and Calico Hills centers of rhyolitic volcanism possess no prominent collapse features, though radial structures caused by doming may be present. The distribution of volcanic structures may be inferred from Figures 17-20.

Some dikes and sills, perhaps intruded along local Basin-Range faults, are associated with basaltic eruptions in the NTS area (Carr, 1974). One intrusive complex at the northwest corner of the Piaute Ridge Quadrangle (Byers and Barnes, 1967) is particularly noteworthy (Figure 25). Many basaltic flows in the area (Figure 22) may be associated with similar shallow assemblages of feeder dikes and sills.

# Cenozoic Extensional Structures

Extensional structures occur where the minimum axial stress is horizontal and the maximum is vertical (Hubbert, 1951). This stress environment creates conditions where lithostatic pressure may exceed the rupture strength of the crust allowing gravity to act as the deforming force. The resulting deformation causes shear fracture, vertical thinning, and crustal extension (Hills, 1963). Faults produced in this manner are called normal, gravity, or Basin and Range faults.

## Normal Faults in the Ranges

The inception of extensional, normal faulting was closely related in time (~20-35 m.y. ago), if not in space, to the initiation of silicic volcanism in the southern Great Basin. Extensional stresses responsible for normal faults may have contributed to the opening of pathways along which volcanic material migrated upward from deep within the crust (Ekren et al., 1968; Crowe, 1978; Zoback and Zoback, 1980). However, the horsts and grabens expressed topographically by the present ranges and basins, respectively, did not form until outbreak of the last pulse of silicic volcanism, about 15-20 m.y. ago (Ekren et al., 1968; Stewart, 1980; Noble, 1972).

According to some investigators inactive normal faults in the NTS area trend northeast and northwest, whereas a younger set, perhaps still active, generally strikes north (Ekren et al., 1968; Wright et al., 1974; Zoback and Zoback, 1980). Others interpret the north trending faults as inactive, with current activity focused along the northeast trending structures (Carr, 1974). In the southeastern portion of the NTS, normal faults generally strike northeast, apparently related to drag folding of original north trending faults along the Las Vegas Valley Shear Zone (Stewart, 1967; Albers, 1967; Carr, 1974). Dips of the normal fault planes range from about 45° to nearly vertical, apparently clustering at about 60°-70° (Hamilton and Myers, 1966; Carr, 1974). Offsets range from inches to thousands of feet (Carr, 1974).

## Normal Faults in the Basins

Drilling and geophysical exploration of Yucca Flat has revealed a complex pattern of normal faults beneath the alluvium. Some faults extend into and even though the alluvium to the basin surface (Carr, 1974). The Yucca Fault, approximately along the axis of Yucca Flat, sharply displaces the alluvium by as much as 10-60 ft. Because the alluvium at the surface is young, this indicates that movement has occurred along the Yucca Fault sometime during the last few thousands to tens of thousands of years. In the subsurface, maximum vertical displacement along the east-dipping Yucca Fault is about 700 ft. The Carpetbag Fault Zone to the west also dips east and exhibits perhaps 2000 ft of displacement since deposition of the Timber Mountain Tuff (Carr, 1974). The Carpetbag Fault is the western boundary of the Yucca Flat graben, whereas the Yucca Fault appears to be the eastern boundary of an intrabasin buried horst. A system of faults between the Yucca and Carpetbag Faults dips west and displays about 2000 ft of vertical displacement along the western edge of the intrabasin horst. Carr (1974) suggests that formation of Yucca and Frenchman basins is relatively recent, with basin formation superimposed on older north-south Basin and Range faults.

Though detailed drill hole information is sparse in other basins of the NTS and Great Basin, geophysical data (Thompson, 1959; Thompson et al., 1967) and

topographic analysis of fault scarps in alluvium (Wallace, 1977) suggest structural complexity and current fault movement occur in basins throughout Nevada. Apparently, the basins are at least as, or more, structurally complex than the ranges. In this context, the basins are perhaps best characterized as topographically low blocks of faulted terrain with irregular, suballuvial structural geometry. Basin and Range faults may occur with almost equal geographical frequency throughout the Great Basin irrespective of Basin or Range topography, though most faults in the basins are obscured by alluvium. Boundaries between the ranges and basins are generally zones of step faults with variable widths rather than single range-bounding faults.

## Folds

Folding is generally not observed along Basin and Range faults, though some large monoclines do occur, such as along portions of the south flank of Skull Mountain. This monocline may be related to either structural or depositional draping over the fault zones. Carr (1974) suggests that minor drag and reverse drag folds may be present adjacent to other normal faults in the NTS area. He also suggests that many of the gravity faults possess a significant component of strike-slip movement.

## Strike-Slip Structures

If the maximum and minimum stress axes are both horizontal, lateral shearing is the stress relief mechanism (Hills, 1963). Because the earth's crust approximates a horizontally layered condition, vertical displacements are much easier to identify and measure than horizontal translations. This is particularly true in complex terrain affected also by vertical displacements, such as southern Nevada. Therefore, evidence for strike-slip movement tends to be indirect.

Both left-lateral and right-lateral shearing has occurred in the NTS vicinity (Figures 27 and 28). Right-lateral movements along the Las Vegas Valley south of the NTS (Longwell, 1960) and the Walker Lane to the north (Locke et al., 1940) have been deduced from two lines of evidence: (1) apparent bending of structural and topographic features and (2) apparent displacement of miogeosynclinal isopach trends (Ross and Longwell, 1964; Longwell, 1960, 1974, Stewart, 1967; Albers, 1967). Ferguson and Muller (1949) suggest that the curving isopach trends in the Walker Lane conform to the outline of ancient shorelines rather than to post-depositional structural trends. Notwithstanding, current thinking favors the strike-slip hypothesis to partly explain the diverse structural and stratigraphic patterns near the Las Vegas Valley and Walker Lane (Carr, 1974).

## Las Vegas Valley-Walker Lane Shear Zone

Assuming originally straight isopachs for the Eureka Quartzite, Ross and Longwell (1964) estimate 25-40 mi of right-lateral offset across the Las Vegas Valley. Longwell (1974) later increased this estimate to 42 mi based on inferred movement of a granite mass relative to the position of local sedimentary deposits derived from the granite. He modified Burchfiel's (1965) estimate of 27 mi by including effects of horizontal bending on either side of the rupture zone. Based on isopach trends in the Sterling and Zabriski quartzites and Wood Canyon Formation, Stewart (1967) concluded that 30 mi of right-lateral offset has occurred along the Las Vegas Shear Zone.

Albers (1967) estimates 80-100 mi of horizontal displacement along the Walker Lane, though total displacement may be distributed among several strike-slip faults and flexures. Named faults along the Walker Lane include the Soda Springs Valley and Cedar Mountain faults northwest of Tonopah (Albers, 1967). Ferguson and Muller (1949) suggest about four miles of displacement across the Soda Springs Valley Fault, and Nielson (1963) estimates about nine to ten miles. Other estimates of displacement across the Walker Lane are scarce though Shawe (1965) agrees with Albers (1967) that movement has been considerable. Gianella and Callaghan (1934) and Shawe (1965) report topographic evidence of right-lateral movement in the Walker Lane associated with an earthquake in 1932.

Volcanic deposits at the NTS obscure the relationship between the Las Vegas Valley Shear Zone and the Walker Lane, though it is reasonable to assume the two zones are related. From southeast to northwest, the Wahmonie-Salyer, Timber Mountain, Black Mountain, and Stonewall Mountain volcanic centers lie on a line connecting these two shear zones. These volcanoes may be genetically related to the same crustal weakness responsible for regional shearing. Continuity of the Walker Lane-Las Vegas Valley shear zones has been proposed by Albers (1967) and Carr (1974) among others, although Suppe et al. (1975) consider the two zones as tectonically different.

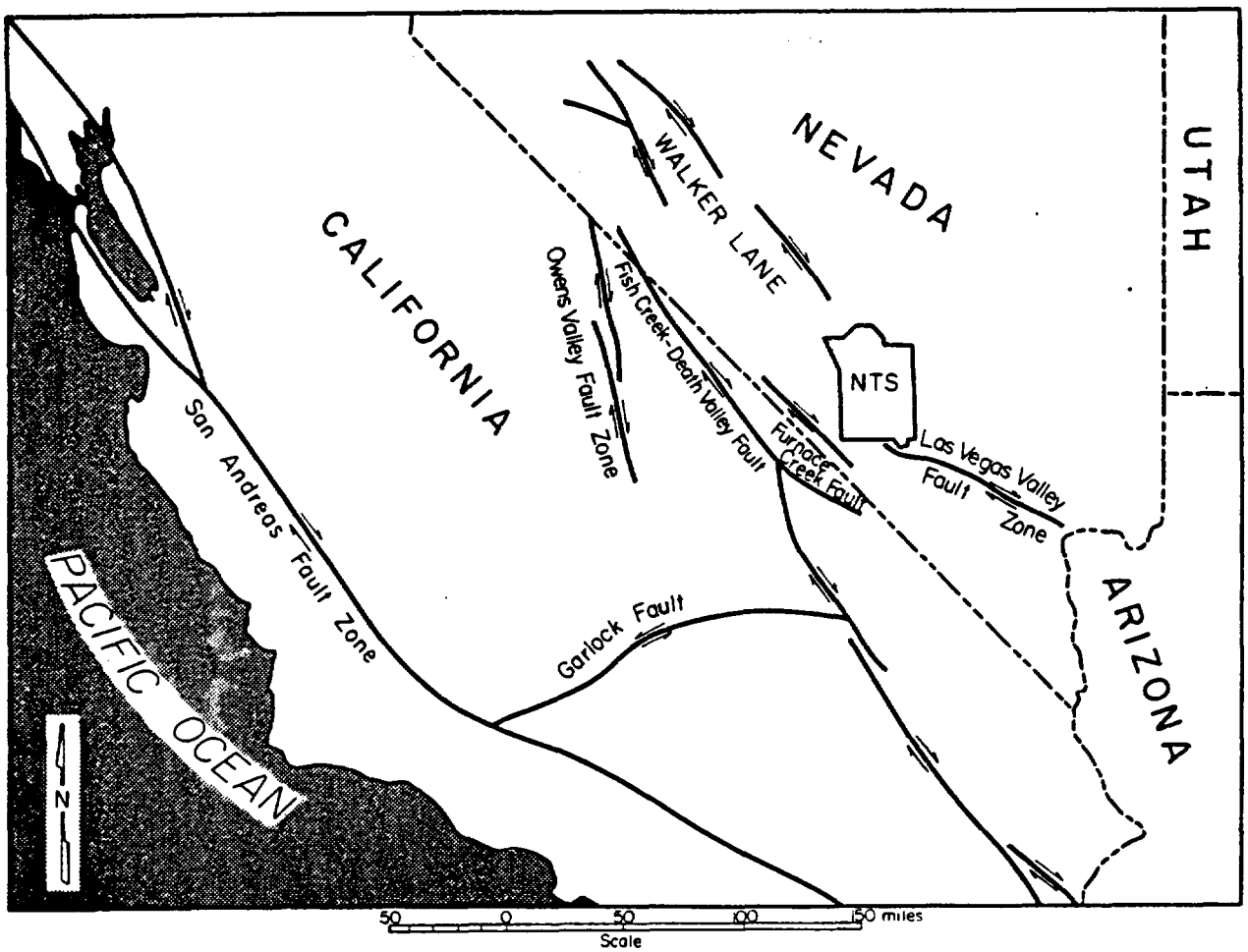


**Figure 28.** Locations of Major Strike-Slip Fault Zones in Southern Nevada and California (from King and Beikman, 1974 and W. J. Carr, Informal Communication)

## Other Great Basin Shear Zones

Large right-lateral shear zones also occur in the California portion of the Great Basin (Figure 28). Bateman (1961) reports as much as 16 ft of right-lateral displacement during a large earthquake in 1872 along the Inyo Fault Zone of Owens Valley. Ross (1962) claims that total post-Jurassic, right-lateral displacement in Owens Valley has been negligible, perhaps a few miles at most. Better agreement is obtained for the combined amount of right-lateral movement along fault zones in Death Valley and Furnace Creek Wash. Estimates vary as follows: less than 10 mi (Wright and Troxell, 1966), about 12 mi (Noble and Wright, 1954); about 20 mi (McKee, 1968); 15-30 mi (Drewes, 1963); and more than 50 mi (Stewart, 1967).

If considered together, right-lateral shear zones of the southern Great Basin comprise a major tectonic element of the western United States. Total lateral shearing may exceed 150-200 mi. If movement along the San Andreas Fault in western California is included, this figure could double.

## Age of Right-Lateral Faulting

Determining when right-lateral movement occurred is even more troublesome than establishing the magnitude of movement. Bending of range topography and fault trends indicates that strike-slip displacement occurred later than formation of the present ranges (Carr, 1974). This assumes that the present ranges were not raised along reactivated, older structures which were bent before reactivation. Ekren et al. (1968) conclude that most right-lateral bending in the southern NTS area occurred during the last 17 m.y. Longwell (1974) places older and younger brackets on movement of the Las Vegas Shear Zone at 17 m.y. and 11 m.y., respectively. Gianella and Callaghan (1934) and Bateman (1961) present evidence for current right-lateral displacement along shear zones northwest of the NTS.

Alternate hypotheses assign most right-lateral movement to much earlier times. Albers (1967) argues that movement along the Walker Lane began in Early Jurassic (about 150 m.y. ago). He notes that Jurassic plutons conform to the structural bends associated with right-lateral movement but notes that the plutons are not internally sheared. He also cites tight folding in the Miocene Esmeralda Formation as evidence for at least minor Miocene to Pliocene shearing along Walker Lane. The same investigator argues for post-early Tertiary movement along the Las Vegas Valley Shear Zone because local Cretaceous and younger rocks are deformed along right-lateral trends. Thus, Albers (1967) suggests that right-lateral movement in southern Nevada started in the Mesozoic and continued at least until the early or middle Tertiary.

Invoking plate tectonic theory Atwater (1970) follows Wise (1963) in suggesting that a diffused plate boundary exists between the Great Basin and San Andreas shear zones. Atwater (1970) assigns all movement to ages younger than 30 m.y. with most as less than 6 m.y., at least along the San Andreas system. This suggestion is not incompatible with the conclusions of Carr (1974) and Longwell (1974) based on local evidence.

## Left-Lateral Faulting

Left-lateral shearing in the NTS area is not so easily related to regional structures. Carr (1974) suggests a zone of northeast trending left-lateral shearing occurs in the southern NTS area with displacement broadly distributed among numerous individual fault planes (Poole et al., 1965b; Orkild, 1968; Ekren and Sargent, 1965; Hinrichs, 1968a). Three major fault zones of left-lateral movement have been identified: the Mine Mountain (not to be confused with the Mine Mountain Thrust Fault); the Cane Springs; and the Rock Valley fault zones (Figures 27 and 29). Another left-lateral shear zone may transect Timber Mountain (Carr, informal communication) (Figure 27).

Carr (1974) argues that left-lateral movement has been slight and is a component of oblique-slip displacement. The left-lateral faults offset, and in turn, are offset by right-lateral movement along the Las Vegas Valley Shear Zone, indicating simultaneous development of the two strike-slip systems (Carr, 1974). Carr (1974) cites stratigraphic evidence of decreasing activity along the left-lateral system beginning about 11 m.y. ago, with only minor offsets continuing until the present. Rogers et al. (1977) present evidence of slight current seismic activity along these left-lateral shear zones.

Carr (1974) suggests that left-lateral faulting is complementary to right-lateral faulting along the Las Vegas Valley. This hypothesis is supported by observations of Longwell (1974) and Anderson (1973) that similar left-lateral shearing is evident near Lake Mead at the southern end of the Las Vegas Valley. If the Walker Lane and Las Vegas Valley are portions of the same right-lateral shear zone, a left-lateral sigmoidal offset of this shear zone may be buried by the volcanic rocks at the NTS. This offset may result from and/or be the cause of the left-lateral movement in the southwestern NTS.

Suppe et al. (1975) suggest that left-lateral faulting in the NTS area may be an insipient transform system separating two major crustal plates. They further suggest that left-lateral faulting in southern Nevada is related to the currently active left-lateral Garlock Fault (Carter, 1971; Davis and Burchfiel, 1973) in California. If correct, this implies increasing movement along the left-lateral faults should be expected, contrary to hypotheses by Carr (1974) and others which suggest decreasing activity.

## Summary of Tectonics in the NTS Area

In summary, the Cenozoic tectonic environment of the NTS region includes a complex suite of fault systems characterized by northeast trending left-lateral faulting, northwest trending right-lateral faulting and associated bending, and northwest, north and northeast trending Basin and Range faulting. These features are all consistent with crustal extension. The Cenozoic structures clearly indicate a dramatic shift in tectonic style from the Mesozoic Era when crustal compression produced thrust faulting and recumbent folding. Carr (1974) suggests that current extension is aligned about N 50° W in the NTS area. The transitional period between compressional and extensional tectonics is poorly preserved in the geological record at the NTS and its duration and character remain largely undefined.

# Miscellaneous Structures

## Joints and Fractures

Many small structures affect the properties of rocks at the NTS. Every rock body is jointed and fractured to some degree. As the surface is approached from depth, joints tend to increase in length, width, and frequency; presumably because structural unloading causes accompanying removal of overburden by erosion, causing, in turn, dilation of the rock mass. Many joints formed in this manner are quasi-parallel to the surface.

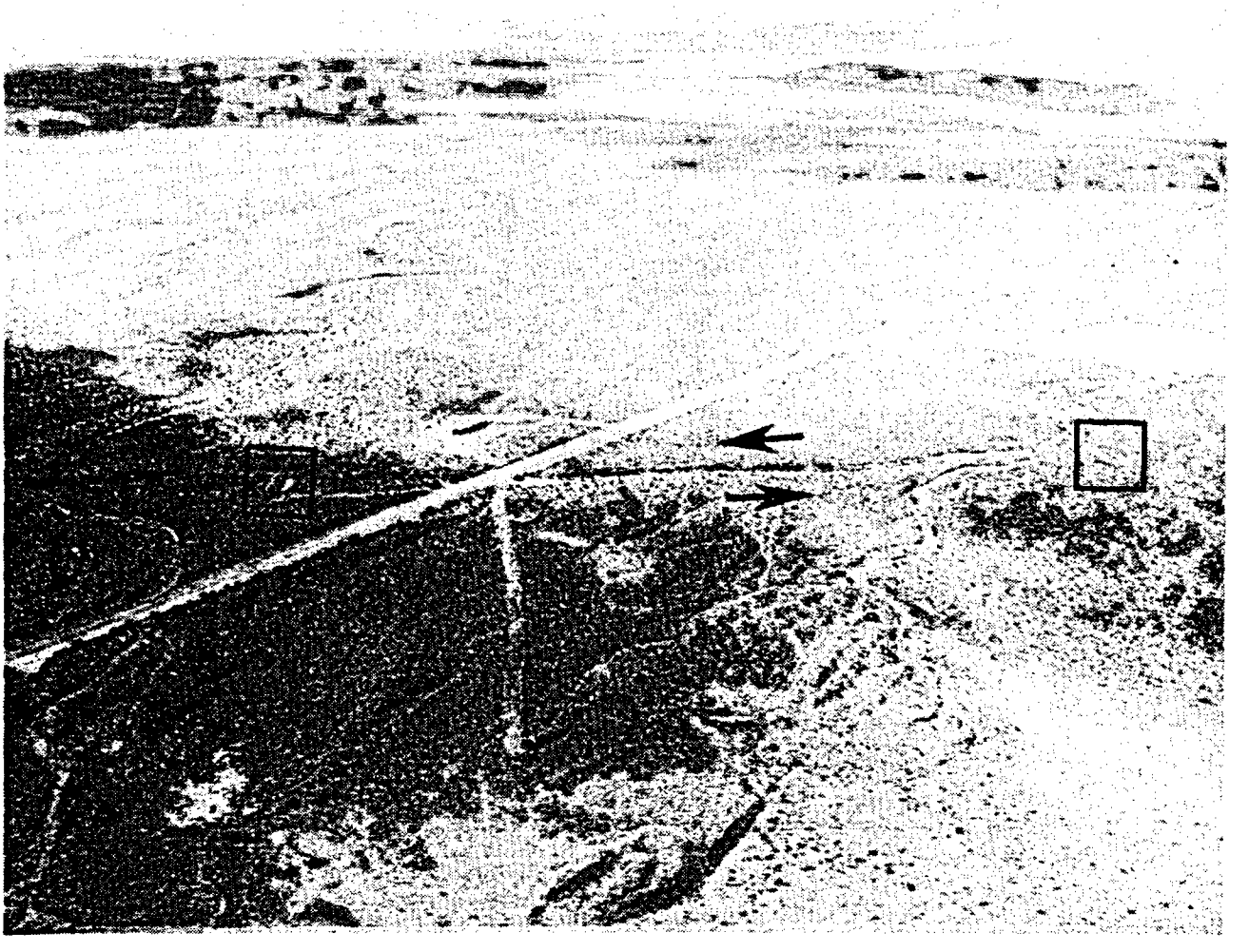

**Figure 29.** Fault Scarp Offsetting Alluvium in Rock Valley, NTS. Presumed movement is at least in part left-lateral (arrows). Note trenches across fault (boxes) to investigate the nature and age of movement. Oblique aerial view is southwest. Photo courtesy of Pan American, Inc.

Horizontal fractures commonly occur along bedding planes in sedimentary rocks. Such fractures, or parting planes, may be inherited from differential loading produced during sedimentation and compaction.

Vertical or inclined joints generally occur in sets with distinctive orientations. They are commonly oriented similar to faults and fold axes, suggesting a genetic relation to stresses responsible for tectonic deformation. Also, earth tides produce recurrent, though locally very small, stresses that may contribute to joint development.

Igneous rocks contract upon cooling and tend to develop joints whose spacing and orientation depend on the thermomechanical properties of the cooling rock mass and its environs. In addition, volcanic rocks, particularly ash flows, may experience devitrification, hydration, and attendant alteration to secondary minerals, all of which involve volume changes that may induce jointing.

Near the surface, all rocks weather. Attendant mineralogical transitions result in volume changes and local stresses. These stresses may accentuate fracturing along weakness planes determined by the rock fabric. Physical changes of state, e.g., freezing of water or precipitation of salts from solution, can expand incipient cracks. Diurnal heating and cooling may contribute to the formation of fractures in the upper few millimeters of exposed rocks, though observations of competent rocks on the moon's surface tend to discount this process. In short, all brittle rocks are fractured to some degree due to a myriad of interrelated processes.

Regional stresses applied to a rock body at depth will form a set of intersecting joint planes and microfractures. Near the surface these initially tight fractures will open, and ground water may enter the joints. Weathering ensues and the cracks expand, setting up new stresses and creating new sets of microfractures. With continued unloading, the process migrates throughout the entire volume of the rock mass. In this manner, existing fractures tend to serve as a focus for stresses that produce new fractures. Fracture intersections compound the rock's vulnerability. Commonly, fracture intersections show signs of rock disintegration even at great depth. After fractures become open to circulating ground water, chemical conditions determine whether solution or precipitation will occur. Many fractures are completely filled with precipitated minerals.

## Orientation and Frequency of Joints and Fractures

The three-dimensional nature of fracture systems can be determined only by exhaustive field mapping, and then only for the surface. Projections to depth are risky. Core samples from drill holes provide local information about fractures at depth, but extrapolation to proximal volumes is tenuous.

However, some generalizations can be made about the nature of fracture systems at the NTS. Regional joint systems were produced by both compressional and extensional stress environments. According to standard rock failure theory (Hills, 1963) compressional fractures should trend northeast, parallel to fold axes and thrust planes, and at about 30° on either side of the principal compression axis. These fractures should tend to dip between about 45° and 70° from horizontal. Extensional fractures should trend about north, northeast, and northwest parallel to Basin and Range and strike-slip faults. These joints should dip about 60°-90°. Fractures produced by volcanic activity tend to be limited to zones of volcanic doming and are radially and concentrically oriented, similar to the volcanic fault systems. Cooling fractures in igneous rocks may occur in any orientation, but they are generally polygonal in plan view and columnar in section. Occasionally, curvi-linear splays of fractures are observed extending across some thickness of individual flow units. Individual cooling fractures are usually limited to portions of individual cooling units, whereas tectonic fractures commonly cross flow-unit boundaries.

In the Ranger Mountains-Banded Mountain area individual fractures are spaced 0.5-2 ft apart and extend unbroken and without direction change for 10's to 100's of feet (P. J. Barosh, written communication in Winograd and Thordarson, 1975). D. L. Hoover and others (written communication in Winograd and Thordarson, 1975) reported joint spacings of less than one foot and up to 43 joints per ft in drill core from the Tippipah Limestone west of Yucca Flat. Lengths of uninterrupted individual fractures range from 0.1 ft to 9 ft, and aperatures vary from less than 0.1 in. to 0.4 in. Maldonado et al. (1979) describe fracture frequencies of about 4 to 5 per ft in core from the Eleana agrillite. Fracture dips generally increase with depth from about 20° to 60° near the surface to about 50° to 80° below 1500 ft depths (Maldonado et al., 1979).

The frequency of cooling fractures in volcanic rocks tends to vary directly with density and brittleness of flow units (Winograd and Thordarson, 1975). Accordingly, welded tuff and basalt tend to be more densely fractured than nonwelded tuff. For different portions of a single flow unit, Winograd and Thordarson (1975) report fracture densities of one fracture per 1/4 in. for an upper vitrophyre, one per 4-8 in. for a central welded tuff, and none in a basal nonwelded flow. Blankennagel and Wier (1973) report joint spacings of 0-7.5 fractures per ft of core derived from welded tuffs at Pahute Mesa, with an average spacing of about 1 fracture per 1-2 ft. For nonwelded tuffs, the average fracture frequency drops to about 1 per every 10 ft (Blankennagel and Wier, 1973). McKeown and Dickey (1961) mapped joint densities of less than 1 per 2-3 ft in nonwelded tuffs of the U12e tunnel complex at Rainier Mesa, and showed fracture free sections up to 40 ft in length. Langkopf and Eshom (1981) mapped intense fracturing in both welded and nonwelded tuffs in the N12g tunnel complex. Spengler et al. (1979a) report average joint frequencies in core from Yucca Mountain of 4.5 to 7.4 per 10 ft intervals of welded Paintbrush tuff, 2.4 to 3.6 per 10 ft intervals of moderately welded Crater Flat Tuff, and only about 1 per 10 ft interval of nonwelded tuffs. Fracture dips in tuffs at Yucca Mountain cluster at about 45° and 80° to 90° depending on the stratigraphic unit.

Sedimentary rocks also exhibit varying fracture frequencies depending on physical properties of the rock. Winograd and Thordarson (1975) generalize that fine-grained carbonates support more fractures (about 1 per in.) than coarse-grained carbonates (about 1 per ft). Quartzites, sandstones, and siltstones tend to intensely fracture perpendicular or oblique to bedding, and individual joints are commonly unconnected and tightly closed. These investigators report that microfractures in quartzite are so dense that almost every grain is bounded by a fracture. Shales, on the other hand, tend to behave in a ductile fashion and seal many of the fractures.

## Giant Playa Fractures

Giant fractures in Yucca Lake deserve special mention (Figure 30). Carr (1974) describes these features as a series of northeast trending fractures which have migrated southward across the surface of Yucca Lake between about 1950 and 1969. These cracks apparently are not the giant, polygonal dessication cracks common to many desert playas (Wilden and Mabey, 1961; Neal et al., 1968). Rather, they appear to be of tectonic origin (Carr, 1974). When open, they probably extend to significant depths, perhaps to rocks beneath the alluvium. This is inferred from the fact that significant quantities of water from Yucca Lake (~300 acre ft), if present, rapidly pour into the cracks when they abruptly open. This tends to fill the cracks with sediment and render them "inactive," whereupon a new crack tends to abruptly form to the southeast at some later time. The origin of these cracks is unknown, but Carr (1974) notes that they all lie along a north trending zone of faulted alluvium that extends from Mercury to Sand Spring Valley. However, no displacements of the playa surface have been noted along the giant cracks. Evidence of other large cracks is found at Groom Lake Playa, Frenchman Flat Playa, Indian Springs Valley Playa, and an unnamed playa about 4 mi northwest of the Papoose Range (Carr, 1974). Many of these, except perhaps in Frenchman Flat, may be the more common variety of giant polygonal dessication cracks (J. T. Neal, oral communication).

## Man-Made Structures

Man-made fractures in the NTS area are produced by four general types of activity: underground explosive testing (predominantly nuclear); hydraulic fracturing; mining; and drilling. Fractures from underground explosions occur at Yucca Flat and Pahute Mesa (Wilmarth and McKeown, 1960; Borg, 1973; Borg et al., 1976; Barosh, 1968). Barosh (1968) mapped concentric and radial fractures extending about 1000 ft from the surface point directly above an explosion. Numerous fractures developed along trends parallel to the master joint sets in rocks adjacent to the explosion site. Fracture patterns around most explosion sites led Barosh (1968) to conclude that much explosion-induced fracturing occurs along natural joint trends. Similarly, fault zones near explosions tend to absorb explosion-induced energy (Barosh, 1968; Dickey, 1968; Carr, 1974). Barosh (1968) describes one explosion that produced movement along a 5000 ft segment of the Yucca Fault, and Dickey (1968) calculated that seismic energy from the fault movement exceeded the seismic energy caused by the explosion. The numerous subsurface explosions in Yucca Flat and Pahute Mesa have undoubtedly had significant effect on fracture patterns and frequency in the areas near the subsurface explosions.

Commonly, rocks collapse into underground cavities produced during explosions, and collapse craters form at the surface. Craters in Yucca Flat range from about 40 ft-1500 ft in diameter (Barosh, 1968). The cavity is commonly connected to the surface crater by a rubble chimney (Borg et al., 1976) which may not extend to the surface depending on the depth of the explosion, explosive energy, geometry of the explosion environment, and character of the surrounding material. Cavity diameters generally range from about 30-120 ft and chimney heights from 30-1500 ft (Borg, 1973). Intense fracturing around the cavity extends about 2-4 times the radial distance from the explosion point to the cavity wall. The cavity commonly contains glassy material intermixed with rock rubble, whereas the rubble chimney is mostly brecciated rock. The cavity, chimney, near cavity fractures, and distal fractures characterize a unique set of man-made structures associated with underground explosions common to the NTS area, predominantly Yucca Flat and Pahute Mesa.

Mining and drilling operations have been extensive on and near the NTS. A complex of tunnels exists beneath Rainier Mesa just west of northern Yucca Flat. Unloading stresses and fractures attendant to mining generally do not propagate more than two tunnel diameters into the host rock (A. R. Lappin, oral communication). The same is probably true for drill holes. Hence, the area of man-induced structures is probably confined to a radius of about 25 ft for the tunnel complex (assuming an average tunnel diameter of about 10 ft) and 1-2 ft for the average drill hole of about 8-12 in. diameter.

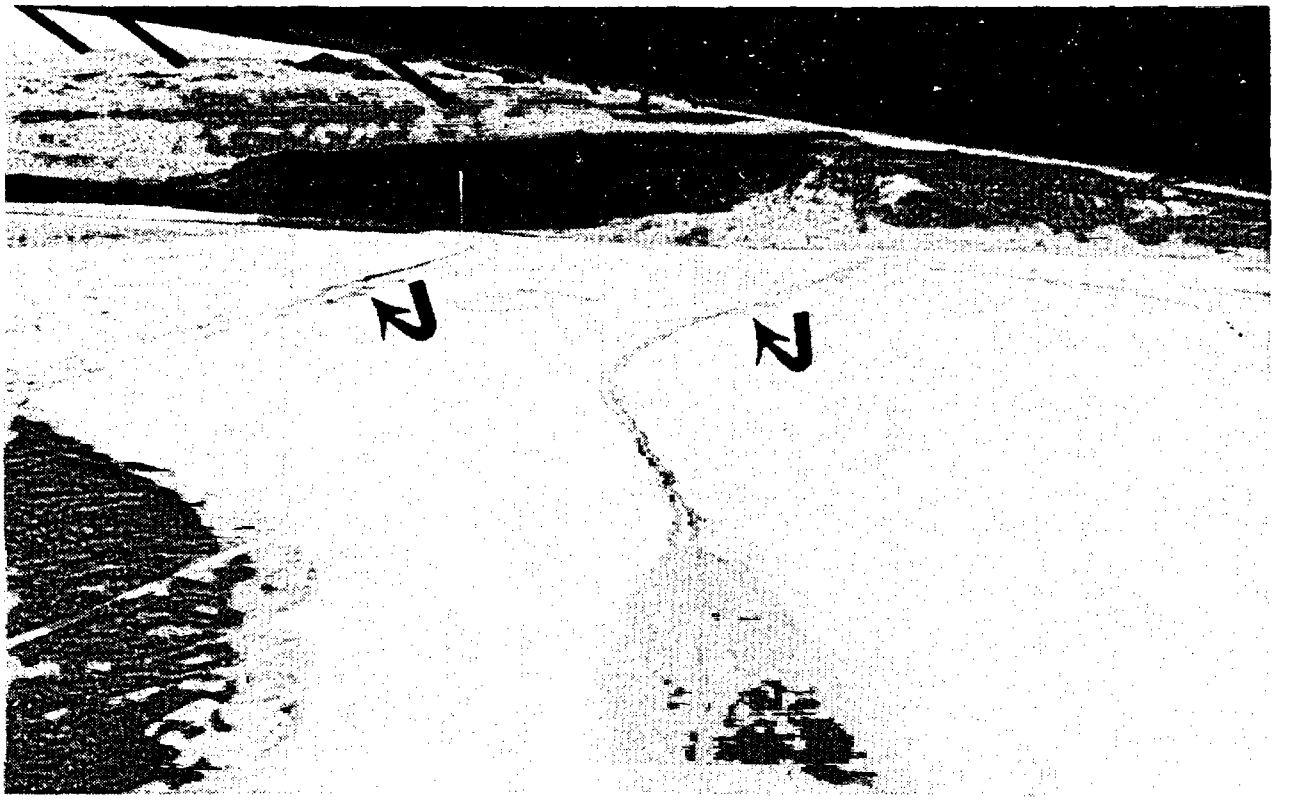

**Figure 30.** Giant Fractures (Arrows) Across the South End of Yucca Flat Playa. Oblique aerial view is southwest. Photo courtesy of Pan American, Inc.

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# APPENDIX

## (Taken directly from Winograd and Thordarson, 1975, pages C117-C119)

Pre-Cambrian and Paleozoic miogeosynclinal rocks, Tertiary volcanic and sedimentary rocks, and Quarternary and Tetiary valley fill are grouped into 10 hydrogeologic units. The grouping is based on similar hydraulic properties, lithologic character, and stratigraphic position. The hydrogeologic units, in order of decreasing age, are: Lower clastic aquitard, lower carbonate aquifer, upper clastic aquitard, upper carbonate aquifer, tuff aquitard, lava-flow aquitard, bedded-tuff aquifer, welded-tuff aquifer, lava-flow aquifer, and valley-fill aquifer. The lower clastic aquitard, the lower carbonate aquifer, and the tuff aquitard control the regional movement of ground water.

The coefficient of transmissibility of the lower clastic aquitard is less than 1,000 gpd per ft. The effective interstitial porosity of 20 cores ranges from 0.6 to 5 percent and has a median value of 1.9 percent. The coefficient of permeability of 18 cores ranges from 0.0000007 to 0.0001 gpd per sq ft and has a median value of 0.000002. Although the clastic strata are highly fractured throughout the study area, regional movement of water through these rocks is probably controlled principally by interstitial permeability rather than fracture transmissibility because (1) the argillaceous strata have a tendency to deform plastically, (2) fractures in the brittle quartzite sequences tend to be sealed by interbedded micaceous partings or argillaceous laminae, and (3) the clastic rocks have low solubility.

The highly fractured to locally brecciated lower carbonate aquifer consists of limestone and dolomite. On the basis of pumping tests of 10 wells and regional flow analysis, the coefficient of transmissibility of the aquifer ranges from 600 to several million gallons per day per foot. Core examination indicates a fracture porosity of a fraction of 1 percent. The effective intercrystalline porosity of 25 cores ranges from 0.0 to 9.0 percent and has a median value of 1.1 percent. The coefficient of permeability of 13 cores ranges from 0.00002 to 0.1 gpd per sq ft and has a median value of 0.00008.

The water-bearing fractures are probably solution-modified joints, fault zones, and breccia. Drill-stem tests in eight holes suggest that the water-bearing fractures are few and widely spaced, are present to depths of at least 1,500 feet beneath the top of the aquifer and as much as 4,200 feet below land surface, do not increase or decrease with depth, are are no more abundant or permeable immediately beneath the Tertiary-pre-Tertiary unconformity than elsewhere in the aquifer.

The lower carbonate aquifer contains several solution caverns in outcrop. One of the caverns, Devils Hole, reportedly extends at least 300 ft vertically into the zone of saturation. The caverns probably do not constitute a hydraulically integrated network of solution openings, except possibly near major discharge areas; variations in fracture transmissibility control the regional movement of ground water through the aquifer.

The upper clastic aquitard consists principally of argillite (about two-thirds of unit) and quartzite (about one-third of unit). Unlike the lower clastic aquitard and the lower carbonate aquifer, which are thousands of feet thick, of relatively uniform lithology, and widely distributed, the upper clastic aquitard is of hydrologic significance only beneath western Yucca Flat and northern Jackass Flats. In much of the area, it is represented by time-equivalent carbonate rocks, has been removed by erosion, or occurs well above the water table. The coefficient of transmissibility of this aquitard is probably less than 500 gpd per ft; interstitial permeability is negligible.

The tuff aquitard consists primarily of nonwelded ash-flow tuff, ash-fall (or bedded) tuff, tuff breccia, tuffaceous sandstone and siltstone, claystone, and freshwater limestone. Despite the widely differing origins of these strata, they generally have one feature in common: their matrices consist principally of zeolite or clay minerals, which are responsible in part for the very low interstitial permeability of these relatively porous rocks. Strata composing the aquitard have

moderate to high interstitial effective porosity (median values ranging from 10 to 39 percent), negligible coefficient of permeability (median values ranging from 0.00006 to 0.006 gpd per sq ft), and very low coefficient of transmissibility (less than 200 gpd per ft). Evidence from several miles of tunnels indicates that the regional movement of ground water through the tuff aquitard is probably controlled by interstitial permeability rather than by fracture transmissibility.

The welded-tuff aquifer consists of moderately to densely welded ash-flow tuff. The coefficient of transmissibility of the aquifer at four well sites ranges from 200 to more than 100,000 gpd per ft and is probably controlled principally by interconnected primary (cooling) and secondary joints; interstitial permeability is negligible.

The valley-fill aquifer consists of alluvial-fan, mudflow, fluvial deposits, and lake beds. The coefficient of transmissibility of the valley-fill aquifer at six well sites ranges from 800 to about 34,000 gpd per ft; average interstitial permeabilities range from 5 to 70 gpd per sq ft.

Owing to the complex structural and erosional history of the area, the subsurface distribution and the saturated thickness of the hydrogeologic units differ from unit to unit and place to place. The structural relief on the pre-Tertiary hydrogeologic units commonly ranges from 2,000 to 6,000 ft within distances of a few miles and locally is as much as 500 ft within distances of 1,000 ft. Thus, the lower carbonate aquifer, which is generally buried and fully saturated at depths of hundreds to thousands of feet below most valley floors, is only partly saturated along flanking ridges. In contrast, in areas where the lower clastic aquitard occurs in structurally high positions, the lower carbonate aquifer either has been largely removed by erosion or occurs entirely, or largely, above the zone of saturation. Also, because of complex pre-early Tertiary deformation and deep erosion, only a fraction of the 15,000-foot aggregate thickness of the carbonate aquifer is usually present in the zone of saturation. In general, because of its great thickness, several thousand feet of the lower carbonate aquifer lies within the zone of saturation beneath most ridges and valleys of the study area.

Vertical displacement, ranging from hundreds to thousands of feet along block faults, affects the subsurface disposition and saturated thicknesses of the tuff aquitard and the welded-tuff and valley-fill aquifers. Beneath valleys that have deep (700-1,900 ft) water tables, the depth to water also affects the saturated thickness of the tuff aquitard and the welded-tuff and valley-fill aquifers. In Yucca Flat, the valley-fill aquifer is saturated only beneath a 10-square mile area where the aquifer thickness exceeds 1,600 feet. Similarly, the welded-tuff aquifer is only partly to fully saturated beneath the central part of that valley; it is unsaturated beneath margins of the valley, even though it is buried at depths of hundreds of feet.

Both intrabasin and interbasin movement of ground water occurs in the region. Intrabasin movement of ground water from welded-tuff and valley-fill aquifers to the lower carbonate aquifer occurs beneath several of the intermontane valleys of the study area. The volume of flow between the Cenozoic hydrogeologic units and the lower carbonate aquifer is usually small, because the aquifers are separated by the thick and widespread tuff aquitard. In Yucca and Frenchman Flats, water leaks downward at the rate less than 100 acre-feet per year in each valley. In east-central Amargosa Desert and on the upgradient side of major hydraulic barriers cutting the lower carbonate aquifer, intrabasin movement is upward from the lower carbonate aquifer into the younger hydrogeologic units. Interbasin movement characterizes flow through the lower carbonate aquifer underlying most of the valleys and ridges of south-central Nevada. Within the Nevada Test Site, water moves south and southwestward beneath Yucca and Frenchman Flats, Mercury Valley, and the east-central Amargosa Desert toward a major spring discharge area, Ash Meadows, in the Amargosa Desert. The hydraulic gradient ranges from 0.3 to 5.9 feet per mile. Interbasin movement through the carbonate rocks is significantly controlled by geologic structure. In the vicinity of major structures, the lower carbonate aquifer is compartmentalized either through its juxaposition against the lower or upper clastic aquitard along major normal or thrust faults, by the occurrence of the lower clastic aquitard in structurally high position along major anticlines, or by gouge developed along major strike-slip faults. The water levels in the lower carbonate aquifer on opposite sides of such structures differ as much as 500 feet in a single valley and as much as 2,000 feet between valleys, although the hydraulic gradient within each aquifer compartment or block is only a few feet per mile.

Hydraulic, geologic, isohyetal, hydrochemical, and isotopic data suggest that the area hydraulically integrated by interbasin water movement in the lower carbonate aquifer is no smaller than 4,500 square miles and includes at least 10 intermontane valleys. This hydrologic system, the Ash Meadows ground-water basin, may, in turn, be hydraulically connected to several intermontane valleys northeast of the study area from which it may receive significant underflow. The principal discharge from the basin, about 17,000 acre-feet annually (about 10,600 gpm) occurs along a

prominent fault-controlled spring line 10 miles long at Ash Meadows. The discharge of individual springs is as much as 2,800 gpm. Underflow beneath the spring line into the central Amargosa Desert is probable, but its magnitude cannot be estimated. Pahrump and Stewart Valleys, proposed as the major source of the spring discharge at Ash Meadows by earlier workers, contribute at most a few percent of the discharge. The major springs in east-central Death Valley (Furnace Creek Wash-Nevares Springs area) are probably fed by interbasin movement of water from central and south-central Amarghosa Desert, but not from Pahrump Valley.

Five hydrochemical facies of ground water in and adjacent to the study area have been distinguished by percentages of major cations and anions. Ground water that has moved only through the lower carbonate aquifer or through valley fill rich in carbonate detritus is a calcium magnesium bicarbonate type. Water that has moved only through rhyolitic tuff or lava-flow terrane, or through valley-fill deposits rich in volcanic detritus, is a sodium potassium bicarbonate type. Water in the lower carbonate aquifer, in areas of downward crossflow from the Cenozoic aquifers and aquitards, is a mixture of these two types and is designated the calcium magnesium sodium bicarbonate type. It is characterized by about equal quantities of the cation pairs calcium plus magnesium and sodium plus potassium. Water in east-central Death Valley, probably a mixture of water of the third type and water from Oasis Valley, is a sodium sulfate bicarbonate type. Shallow ground water, such as that beneath saturated playas, is informally designated as the playa type. The chemistry of this water (dissolved-solids content as high as 50,000 mg/l) varies widely and depends in part on the depth of the sampling point. Major inferences pertinent to the ground water regimen, made on the basis of hydrochemical data, are as follows:

1. Ground water beneath the Nevada Test Site moves towards the Ash Meadows area.
2. Chemistry of water in the lower carbonate aquifer may not change markedly to depths as great as 10,000 feet.
3. Leakage of water from the tuff aquitard into the lower carbonate aquifer is probably less than 5 percent of the spring discharge at Ash Meadows.
4. Underflow into the Ash Meadows basin, from Pahranagat Valley, may amount to as much as 35 percent of the spring discharge as Ash Meadows.

The estimated velocity of ground water moving vertically through the tuff aquitard into the lower carbonate aquifer in Yucca Flat ranges from 0.0005 to 0.2 foot per year; values toward the lower end of the range are more probable.

The estimated velocity of water in the lower carbonate aquifer beneath central Yucca Flat ranges from 0.02 to 2.0 feet per day. Velocity in the carbonate aquifer berneath the Specter Range ranges from 2 to 200 feet per day. The spread of two orders of magnitude in estimated velocities in the carbonate aquifer in each area is due principally to uncertainty about the fracture porosity of the aquifer.

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