Geologic and Geophysical Maps of the Las Vegas 30′ × 60′ Quadrangle, Clark and Nye Counties, Nevada, and Inyo County, California

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Chapter A
Geology of the Las Vegas 30′ × 60′ Quadrangle

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Introduction

This report of the Las Vegas 30′ × 60′ quadrangle was completed by the U.S. Geological Survey's Las Vegas Urban Corridor Project, National Cooperative Geologic Mapping Program. As one of the fastest growing cities in the country, Las Vegas, Nevada, is challenged by major issues such as water supply and contamination, land subsidence due to ground-water withdrawal, and seismic and flood hazards. The report was designed to respond to the rapid urbanization experienced by Las Vegas by providing earth science information that can be used to investigate these land-use issues.

This report contains two parts: the first, chapter A, describes the geology of the map area and includes the geologic map of the Las Vegas quadrangle, cross sections, and text with description of map units and discussion of the stratigraphic and structural framework. The second part, chapter B, describes the geophysics of the map area and includes isostatic gravity, aeromagnetic, and gravity inversion maps, geophysical cross sections of Las Vegas and Pahrump Valleys, and text describing the geophysical framework. The integration of geologic and geophysical information contained in the report offers a unique opportunity to apply basic geoscience techniques to help solve urban growth issues in this arid population center.

The Las Vegas quadrangle is in the Basin and Range Province, in the southern part of the North American Cordillera. The major valleys in the quadrangle are Las Vegas and Pahrump Valleys; both contain thick Quaternary and Tertiary basin-fill deposits derived from the Spring Mountains and Las Vegas and Sheep Ranges. The Spring Mountains are in much of the western and central parts of the quadrangle, and are the highest mountains in southern Nevada. The southern Sheep and Las Vegas Ranges define the northern margin of Las Vegas Valley.

Quaternary deposits cover about half the quadrangle, underlying most of the urbanized areas. Deposits consist of large coalescing fans that grade downslope to extensive areas of fine-grained sediment indicative of Pleistocene ground-water paleodischarge (Haynes, 1967; Quade, 1986). In the central parts of Las Vegas and Pahrump Valleys, the association of Quaternary fault scarps with these sediments suggests a genetic relationship. The synthesis of new mapping of Quaternary geology for this report illustrates the distribution and recurrence of ground-water discharge in Las Vegas and Pahrump Valleys and provides clues for understanding past and present-day ground-water flow paths and water resources.

Rocks in the Spring Mountains and Las Vegas and Sheep Ranges are mostly Paleozoic marine strata that formed in the transition between the Cordilleran miogeocline and the craton. The distribution of Paleozoic rocks is important because they are part of the Nevada carbonate rock province (Dettinger and others, 1995), and are major ground-water aquifers in Nevada. These rocks are deformed by east-directed Mesozoic thrust faults. The Keystone fault in the eastern Spring Mountains is the easternmost thrust of the Cordillera foreland thrust belt in the region, and is one of the best exposed thrust faults in the entire Cordillera.

The Las Vegas Valley shear zone, which offsets Paleozoic strata and Mesozoic thrust faults in the Spring Mountains and Las Vegas Range, has about 50 km of right-lateral horizontal (strike-slip) displacement. This west-northwest-striking shear zone is concealed beneath thick basin-fill deposits of Las Vegas Valley, and it played a significant role in the tectonic development of Las Vegas Valley. Similarly, the development of northwest-striking Pahrump Valley is closely associated with movement along the State Line fault zone. Both fault zones are right-lateral strike-slip fault zones that formed during episodes of extensional faulting in Cenozoic time. In addition, the Las Vegas Valley shear zone has been termed a transverse zone, oriented nearly parallel to the extension

direction, and accommodating different rates, amounts, and types of extensional strain north and south of it (Rowley, 1998).

Along the eastern Spring Mountains is Spring Mountain State Park, managed by the State of Nevada, and Red Rock Canyon National Conservation Area, managed by the U.S. Bureau of Land Management (BLM). These are two of the region's most popular parks. The quadrangle also includes part of the Desert National Wildlife Refuge in the southern Sheep and Las Vegas Ranges, managed by the U.S. Fish and Wildlife Service. The National Park Service Lake Mead National Recreation Area is in the adjacent Lake Mead 1:100,000-scale quadrangle. This report provides information for Las Vegas visitors about how the rocks and landscape of the area's scenic parks and lands formed.

This report provides an integrated geologic and geophysical framework for ground-water investigations in the Las Vegas region. Geologic and geophysical mapping conducted for this report helped to identify faults, fractures, and other permeable zones that are pathways for ground-water flow, and thus it helps to determine the distribution of potential aquifers and ground-water flow paths and barriers. Several important ground-water studies were completed in conjunction with our mapping, including evaluation of the Las Vegas Valley shear zone for ground-water resource potential (Langenheim and others, 1997, 1998) and assessment of western Las Vegas Valley as a potential artificial recharge site (Langenheim and Jachens, 1996).

Methods

Bedrock map units were compiled from existing published geologic maps in the quadrangle (fig. 1) (Axen, 1985; Bohannon and Morris, 1983; Bingler, 1977; Burchfiel and others, 1974; Carr, 1992; Carr and McDonnell-Canan, 1992; Carr and others, 2000; Ebanks, 1965; Lundstrom and others, 1998; Maldonado and Schmidt, 1991; McDonnell-Canan and others, 2000; Matti and others, 1993; Page and others, 1998), with some reinterpretations and modifications following field checking. The geologic map of Clark County, Nevada (Longwell and others, 1965), was an important source of information used in our compilation. New geologic mapping of bedrock units was completed in the Las Vegas Range, Blue Diamond area, and the Cold Creek area in the northwestern Spring Mountains (fig. 1; W.R. Page, new mapping, this report). Several sets of vertical aerial photographs considerably aided our new mapping, including 1977, 1978, and 1980 U.S. Bureau of Land Management natural color photographs at 1:30,000 scale, and 1973 and 1980 U.S. Geological Survey black and white photographs at 1:30,000 scale.

The age of many Paleozoic units in the Las Vegas quadrangle is based on conodont biostratigraphy. More than 100 new conodont age determinations were added to the many determinations previously published for the study area (Miller and Zilinsky, 1981; Stevens and others, 1996, among others).

Systematic collections for conodonts were taken in measured sections through lower and middle Paleozoic successions in the major thrust plates in order to characterize changes in stratigraphy and to correlate Paleozoic units across thrust plates (fig. 2; Harris and others, in press). These collections improved correlations locally and regionally by allowing more precise dating of stratigraphic units and unconformities as well as permitting the recognition of units that were misidentified or overlooked in previous studies.

Quaternary map units were defined and delineated on the basis of lithostratigraphy, soil development (U.S. Soil Conservation Service, 1985; Sowers and others, 1988; Reheis and others, 1992), geochronology, surface characteristics, and photogeologic characteristics, calibrated by iterative field checking during 1996-99. Several sets of vertical aerial photographs were used, including 1977, 1978, and 1980 U.S. Bureau of Land Management natural color photographs (1:30,000 scale), which were the most useful and main set used; 1973 and 1980 U.S. Geological Survey black and white photographs (1:30,000 scale); and U.S. Geological Survey (1:40,000 and 1:80,000 scale) black and white and false-color IR photographs. Stereoscopic delineation of features was registered to 1:24,000-scale base maps using a PG-2 plotter and by use of orthophotos. Previous Quaternary geologic mapping and fault studies were adapted for this map (fig. 1: Haynes, 1967; Bingler, 1977; Matti and Bachuber, 1985; Matti and others, 1987; Matti and others, 1993; Quade, 1983; Sowers, 1986; Bell, 1981; Bell and others, 1998, 1999; Weide, 1982; McDonnell-Canan, 1989; Dohrenwend and others, 1991; Hoffard, 1991; Maldonado and Schmidt, 1991; Anderson and O'Connell, 1993; Anderson and others, 1995a, b; dePolo, 1998; dePolo and others, 1999; Amelung and others, 1999). These studies were adapted through additional field checking, photointerpretation, and map registration where needed. Samples for thermoluminescence (methods of Millard and Maat, 1994) and U-series dating (methods of Ludwig and Paces, 2002) were collected and analyzed from sites within the map area (tables 1 and 2).

Description of Map Units

Quaternary and Tertiary Rocks

[Pedogenic carbonate stages from Gile and others, 1966]

Gravelly Alluvium, Predominantly of Alluvial Fans

Coarse gravelly alluvium is subdivided below on the basis of age-related characteristics. Provenance accounts for some variability in alluvial fan characteristics that is independent of age. Within the map area, limestone and dolostone dominate the lithology of gravel clasts in alluvial fans derived from much of the Spring Mountains, Las Vegas Range, and Sheep Range. However, different fan compositional characteristics exist where other rock types comprise significant areas of exposed bedrock source area. In the areas flanking

Table 1. Thermoluminescence data.

Sample	Latitude degrees°, minutes', seconds"	Longitude degrees°, minutes', seconds"	Depth (cm)	Stratigraphic position	Modal grain size	% moisture field	% moisture saturated	Dose rate (grays/ky) field	Dose rate (grays/ky) saturated	ED (1) (grays)	TL age (ka) minimum	TL age (ka) maximum
LV-1	36°16'02"	115°03'44"	300	in Qso	silt	4	74	2.3	1.3	519	226	399
LV-2	36°18'22"	115°06'52"	120	basal Qay	fine sand	0.2	27	2.7	2.1	44.2	16	21
LV-3	36°17'07.7"	115°07'46.5"	200	in Qsab	fine sand	3	21	3.85	3.18	382	99	120
LV-6	36°04'53.7"	115°03'48.3"	800	in Qfw	fine sand	0.4	42	3.6	2.2	834	232	379
LV-13	36°22'50"	115°05'44"	200	in Qai	fine sand	3	21	3.85	3.18	382	99	120
LV-14	36°20'54.3"	115°17'15.1"	270	in Qse	silt	0.9	49	4.6	2.9	49	11	17
LV-15	36°20'54.3"	115°17'15.1"	260	in Qse	silt	0.2	42	4.6	3	44	10	15
LV-16	36°20'58.8"	115°17'08.7"	400	in Qsc	silt	1	40	3.6	2.4	86	24	36
LV-17	36°19'06.7"	115°11'46.3"	280	in Qsd	silt	1.4	57	6.7	4	162	24	41
LV-18	36°19'09.6"	115°11'54.5"	350	in Qsc	silt	3.7	50	4.3	2.7	150	35	56
LV-19	36°19'08.9"	115°11'27.9"	600	in Qsa	silt	2	64	4.8	2.8	630	131	225
LV-20	36°19'08.9"	115°11'27.9"	350	in Qsb	silt	2	51	4.5	2.8	402	89	144
LV-21	36°07'10.6"	115°53'15.5"	240	basal Qsc	silt	13	42	5.3	4.1	204	38	50
LV-22	36°07'10.6"	115°53'15.5"	150	in Qsc	silt	6	37	6.4	4.7	132.4	21	28
LV-23	36°07'00.8"	115°53'04.7"	120	in Qse	silt	2.2	27	3.7	2.9	74.5	20	26
LV-24	36°04'28"	115°56'23"	220	in Qd	fine sand	2.6	24	4.3	3.43	8.53	2.0	2.5
LV-25	36°03'03"	115°54'53"	800	in Qbo	silt	5.4	38	2.53	1.87	750.2	297	401
LV-26	36°03'03"	115°54'53"	500	in Qbo	silt	5.6	33	2.78	2.13	584.36	210	274
LV-27	36°02'00"	115°53'18"	200	in Qby	silt	5.2	37	2.92	2.17	616.5	211	284
LV-28A	36°08'05.9"	115°23'58.5"	500	in Qao	fine sand	0.2	44	1.2	0.79	562	468	711
LV-28B	36°08'05.9"	115°23'58.5"	500	in Qao	fine sand	0.2	44	1.2	0.79	468	390	592
LV-29	36°08'06"	115°23'58.5"	500	in Qayy	fine sand	2.6	24	4.3	3.43	8.53	2.0	2.5

Note: ED (1) is equivalent dose at peak temperature T of glow curves

Table 2. Uranium series dates.

Sample	Latitude degrees°, minutes', seconds"	Longitude degrees°, minutes', seconds"	Unit	Depth (m)	U ppm	Th ppm	$^{234}U/^{238}U$	$^{230}Th/^{238}U$	$^{230}Th/^{232}Th$	Age (ka)	Initial $^{234}U/^{238}U$
TSQ2C-A	36°20'57"	115°17'18"	Qsd	surface	1.88	2.05	2.144±10.9	1.0477±9.8	2.87	68 ± 13	2.388±0.25
TSQ2C-B	36°20'57"	115°17'18"	Qsd	surface	3.12	1.82	2.5242+5.45	0.52816+11.7	3.11	24.94 + 4.4	2.636+0.13
WM1A-A	36°05'03"	115°04'00"	Qao	vein at 2m	1.1	0.211	1.7138+1.33	1.89+1.69	30.4	363 + 30	2.996+0.150
WM1A-B	36°04'58"	115°03'52"	Qao	vein at 2m	1.6	0.3571	1.657+1.49	1.82+1.74	24.1	317 + 19	2.712+0.088
MSQ1A-A	36°07'13"	115°53'13"	Qsc	2	3.13	3.12	2.5421+10.7	1.055+8.44	3.17	55 + 8.9	2.802+0.28
MSQ1C-A	36°07'13"	115°53'13"	Qsd	0.3	1.33	2.66	2.3031+30.6	0.74068+40.8	1.33	40.97 + 29	2.464+0.7
MSQ2B-A	36°07'13"	115°53'13"	Qsd	surface	3.33	2.89	2.315+8.3	0.61701+14.9	2.46	32.72 + 7.8	2.443+0.19
MSQ3A-D	36°07'05"	115°53'02"	Qai	2	1.1	0.94	1.6712+6.41	0.96469+7.71	3.45	87.22 + 13	1.859+0.12
MSQ4B	36°05'04"	115°59'16"	Qbw	1	0.86	0.06	1.4249+0.773	1.523+1.37	65.47	370.7+ 43	2.215+0.11

the southern part of the Spring Mountains, the fans of Lovell Wash, Red Rock Wash, and Tropicana Wash include a greater proportion of sandstone clasts, matrix sand, and interbedded sand derived from Lower Permian redbeds and the Jurassic Aztec Sandstone. In contrast, fans derived from the northwest Spring Mountains are dominated by resistant angular quartzite clasts. These fans are generally more bouldery because the quartzite does not weather into pebble-size clasts as readily as do the Paleozoic and Mesozoic carbonates and sandstones. Desert varnish also more readily develops on quartzite clasts. In the southeast corner of the map area, fans derived from the Tertiary volcanics of the McCullough Range also have abundant boulders and have well-developed rock varnish on older fans. The relatively high degree of pedogenic carbonate development in the older fan units in this area may relate, at least partly, to the more resistant nature of these bouldery gravels.

The coarse alluvial gravel units described below were deposited predominantly but not exclusively as alluvial fans. Gravelly alluvium within alluvial fans grades valleyward into pediment veneers, terraces, and washes that overlie or are inset within older fine-grained deposits. Fan gravel at the distal toes of alluvial fans is also commonly interbedded with, and a gradational facies equivalent to, correlative fine-grained deposits. For any given fan or wash, there generally is a decrease in maximum and average clast size away from upland source areas. This affects the size of the original depositional bar-and-swale morphology and how rapidly such morphology is modified by surficial processes and pedogenesis.

For all alluvial map units, gravel and sand are interbedded and intermixed, poorly to moderately well sorted, massive to well bedded, and clast supported to matrix supported. Gravel is angular to subrounded, ranging in size from granules to boulders; maximum clast size decreases with increasing distance from upland source areas. Individual unit descriptions below emphasize surface characteristics, associated soil development, and geochronology that distinguish each unit.

Qay   Young fan alluvium (Holocene and latest Pleistocene)—Noncemented alluvial-fan gravel and sand with weakly developed soil. Includes Holocene and locally latest Pleistocene (younger than about 15 ka) deposits between modern channels, and deposits in numerous modern channels that are too narrow (less than 30 m) to map separately. Etching on surficial limestone clasts ranges from absent on deposits of modern channels to incipient and sparse to moderately developed and common on deposits between modern channels. Bar-and-swale depositional morphology ranges from prominent in modern channels to variably modified and muted by addition of eolian sediment in areas between modern channels. Even in the most muted cases, cobbles and boulders protrude from eolian sand cover, and

relict depositional microrelief is evident on aerial photographs. Desert pavement ranges from absent on deposits in modern channels to loosely packed and weakly developed (especially in areas of relatively low dust-flux, as on the upper part of a fan) to moderately well packed in areas of higher dust flux. Rock varnish, which does not form on most limestone clasts, is generally weakly developed to absent on more siliceous rock types (including siliceous carbonates) except for relict rock varnish not abraded during transport. Typical weak soil development is characterized by a cambic Bw horizon, by the presence of non-cemented stage I-II secondary carbonate morphology (mostly thin coats on clast undersides), and by a gradual increase of sand toward the surface through the top 0.5 m of the deposit. The surficial sand component is considered to be pedogenically mixed and infiltrated eolian sand deposited after fluvial deposition of fan gravel. Age control by radiocarbon and thermoluminescence indicates that **Qay** is predominantly Holocene, and locally as old as 14 ka. Near basin centers, **Qay** either overlies or is inset within fine-grained deposits with abundant radiocarbon dates ranging from about 8 to 12 ky B.P. (Haynes, 1967; Quade, 1986; Quade and others, 1995; Bell and others, 1998, 1999), so it is likely that most alluvium at the surface of **Qay** is Holocene. **Qay** includes deposits correlative to youngest alluvium (**Qayy**) and older deposits (**Qayo**). Minimum thickness of **Qay** ranges from less than 1 m to at least 3 m, as exposed in borrow pits; base of unit is generally not exposed

Qayy   Youngest alluvium (Holocene)—Noncemented alluvial-fan gravel and sand of intermittently active wash complexes. **Qayy** is delineated separately from **Qay** only where there is a markedly greater proportion of modern channels with minimal development of surface etching, varnish, pavement, and soil. Includes deposits of modern channels that are too narrow (less than 30 m) to map separately, as well as Holocene deposits between modern channels. Etching on surficial limestone clasts ranges from absent on deposits of modern channels to incipient and sparse on deposits between modern channels. Bar-and-swale depositional morphology ranges from prominent in modern channels to variably modified and muted by addition of eolian sediment in areas between

modern channels. Desert pavement ranges from absent on deposits of modern channels to loosely packed and weakly developed on deposits between active channels. Rock varnish, which does not form on most limestone clasts, is generally very weakly developed to absent on more siliceous rock types (including siliceous carbonates) except for relict rock varnish not abraded during transport. Typical non-cemented, weak soil development is characterized by the presence of stage I–II secondary carbonate morphology (mostly thin coats on clast undersides), and by a gradual increase of sand toward the surface in the upper 10–30 cm of the unit. The surficial sand component is considered to be a pedogenically mixed and infiltrated eolian sediment deposited after fluvial deposition of fan gravel. Spaulding and Quade (1996) reported four radiocarbon dates ranging from 1070 to 3310 years B.P. (analytical error 40–60 years) on bristlecone pine logs in and on debris flow deposits of Qayy deposits in the Yucca Forest area, northern Gass Peak quadrangle, (included on this map as **Qay**). This unit grades into partly correlative fine-grained facies mapped as **Qfy**. Minimum thickness is 1–2 m; base of unit is generally not exposed

Qayo   **Older young alluvium (Holocene and latest Pleistocene)—**Noncemented gravel and sand with weakly developed soil of alluvial-fan remnants. Associated with surfaces that are 2 m to greater than 10 m above modern channels. Etching on surficial limestone clasts is moderately developed and common. Surface is generally smooth, with general absence of depositional bar and swale morphology. Desert pavement is moderately packed. Rock varnish, which does not form on limestone clasts (Liu and others, 2000), is moderately developed on more siliceous rock types (including siliceous carbonates). Typical non-cemented and weak soil development is characterized by a cambic Bw or incipient Btj horizon, by the presence of stage I–II secondary carbonate morphology (mostly thin coats on clast undersides), and by a gradual increase of sand toward the surface through the top 0.5 m of the deposit. The surficial sand component is considered to be pedogenically mixed and infiltrated eolian sediment deposited after fluvial deposition of fan gravel. Unit locally overlies **Qse**, dated at about 10–16 ka (described below). **Qayo** includes unit E$_2$ of Haynes (1967) and Q$_3$ fan

units of Bell and others (1998, 1999) and Liu and others (2000). Minimum thickness of **Qayo** ranges from less than 1 m to at least 3 m, as exposed in borrow pits

Qaiy   **Younger intermediate fan alluvium (late Pleistocene)—**Cemented alluvial-fan gravel, with interbedded sand; poorly to moderately well sorted; massive to well bedded; clast-supported to matrix-supported. Gravel is angular to sub-rounded, ranging in size from granules to boulders. Surface is characterized by a moderately to tightly packed desert pavement. Though nonsiliceous limestone clasts do not have rock varnish, siliceous clasts possess dark varnish, which impart a darker tone to this unit on aerial photographs. This surface morphology is associated with a soil that typically includes a weakly to moderately developed reddish-brown argillic (Bt) horizon and cemented stage II–IV carbonate morphology. Within the soil profile, the upward decrease in proportion of gravel is due to the addition of eolian material concurrent with pedogenesis. Within and beneath the zone of maximum pedogenic carbonate development, limestone gravel generally is distinctly more cemented than in younger Holocene alluvium. Depositional microrelief is minimal relative to other alluvial units; bar-and-swale morphology is generally absent, and surfaces are smooth between limited areas of erosional dissection. Where distinguished from unit **Qai**, which has similar soil development and surface characteristics, **Qaiy** consists of surfaces that are inset relative to **Qai**, especially within the incised fanhead areas of major washes that provide drainage from relatively high areas of the Spring Mountains, including Lovell Wash, Harris Wash, and Willow Wash. Unit **Qaiy** is not dated, but is hypothesized to be correlative to Lundstrom and others' (1999) unit **Qfiy** on Fortymile Wash near Yucca Mountain, which they dated between about 25 and 50 ka, and to unit **Qscd** on this map which we dated between about 25 to 40 ka. Where mapped separately, unit **Qaiy** is a strath terrace generally less than 2 m thick but grades basinward to thicker alluvial fan gravels included in unit **Qai**

Qai   **Intermediate fan alluvium (late and middle? Pleistocene)—**Cemented alluvial-fan gravel, with interbedded sand; poorly to moderately well sorted; massive to well bedded; clast-supported to matrix-supported. Gravel

is angular to sub-rounded, ranging in size from granules to boulders. Includes modern channels too narrow to map separately. Surface between modern channels is somewhat erosionally rounded where it is more than about a meter above grade of adjoining channels. Surface between modern channels is characterized by a moderately to tightly packed desert pavement and smooth surface, which generally lacks bar and swale depositional morphology. Though nonsiliceous limestone clasts do not have rock varnish, siliceous clasts possess dark varnish, which impart a darker tone to unit **Qai** relative to younger units. This surface morphology is associated with a soil that typically includes a reddish-brown argillic (Bt) horizon and cemented stage II–IV carbonate morphology. Within the soil profile, the upward decrease in proportion of gravel is due to the addition of eolian material concurrent with pedogenesis. Within and beneath the zone of maximum pedogenic carbonate development, limestone gravel generally is distinctly more cemented than in younger Holocene alluvium. Depositional microrelief is minimal relative to other alluvial units; bar-and-swale morphology is generally absent, and surfaces are smooth between limited areas of erosional dissection. At site LV-13 (SW ¼ sec. 19, T. 18 S., R. 62 E.), the eolian part of a buried soil at about 2 m depth within unit **Qai** yielded a thermoluminescence date of about 100-120 ka (table 1). We estimate **Qai** to have been deposited between about 50 ka and 130 ka (the early part of the late Pleistocene, including the last interglacial). Exposed minimum thickness of **Qai** ranges from less than 1 m to at least 5 m; base of unit not exposed

Qau   **Undivided young and intermediate alluvium (Holocene and late Pleistocene)—** Cemented and noncemented alluvial-fan gravel, with interbedded sand; poorly to moderately well sorted; massive to well bedded; clast-supported to matrix-supported. Gravel is angular to subrounded, ranging in size from granules to boulders. **Qau** represents areas in which young fan alluvium (**Qay**), including common intermittent active channels, exists as discontinuous but common (30–50 percent of **Qau**), thin (<1m) veneers over intermediate fan alluvium (**Qai**) in patches that are too narrow to map separately. In the area of the State

Line fault zone to the northeast of Pahrump Valley playa in the southwest part of the map area, **Qau** refers to undivided Quaternary alluvium reworked from **QTa** and **Qbw** (described below)

Qao   **Old alluvium (Pleistocene)—**Cemented gravel and sand of partly eroded alluvial-fan remnants and terraces with well-developed, but degradational soil. The erosional upper surface consists of broadly rounded, accordant ridges that are gradational basinward to less deeply dissected surfaces. Depositional microrelief is absent. The generally well-packed surface pavement that developed on this erosional surface includes variable but often darkly varnished clasts and abundant clasts of pedogenic calcite and calcite-cemented gravel; these clasts impart a lighter surface tone to this unit than to adjoining surfaces of younger map units. Soil typically includes a laminar (stage IV, Gile and others, 1966) but partially eroded K horizon at least 0.5 m thick; an overlying argillic horizon of variable thickness and expression is commonly present. The surface characteristics indicate that the original upper depositional surface of **Qao** has been partially eroded. The pavement, and at least some of the soil, clearly were formed after the latest erosional episode. Accordancy of ridge tops of **Qao** indicates a common single grade in an area that the ridge tops define; this grade is inferred to approximate the original depositional top of **Qao** before erosion. This morphology contrasts with the non-accordant ridge tops of **QTa** (described below), indicating that no original depositional top of **QTa** remains. The accordant surface defined by **Qao** ridge tops is inset within **QTa** where these units are juxtaposed. Where **Qao** adjoins **Qai**, the upper part of **Qao** is clearly at a higher grade and more eroded than **Qai**. **Qao** is not dated. Because the accordant ridge tops of **Qao** can be traced to areas within a few meters of the grade of **Qai**, which has yielded late Pleistocene dates, we infer that **Qao** includes middle and late Pleistocene deposits. Minimum exposed thickness of **Qao** is at least 4 m; base of unit is generally not exposed

Qaoi   **Undivided Pleistocene alluvium—**Cemented sandy gravel and interbedded sand of alluvial fans; consists of numerous inset to superposed veneers of intermediate fan alluvium (**Qai**) on old alluvium (**Qao**), such that surfaces

of both units are interspersed at a scale too small to map separately

QTa   **Gravelly basin-fill alluvium (Pleistocene to late Miocene?)**—Generally well-cemented, sandy gravel and interbedded finer deposits of undivided basin fill. Poorly to moderately well sorted; massive to well bedded; clast-supported to matrix-supported. Gravel is angular to subrounded, ranging in size from granules to boulders. The erosional upper surface consists of rounded, non-accordant ridges between intervening deep gullies. Depositional microrelief is absent. Variably well-packed surface pavement includes abundant clasts of pedogenic calcite and calcite-cemented gravel; these clasts impart a lighter surface tone to this unit than is present on adjoining surfaces of younger map units. Soil typically includes a laminar (stage IV, Gile and others, 1966) but partially eroded K horizon at least 0.5 m thick; an overlying argillic horizon of variable thickness and expression is commonly present. The surface characteristics indicate that the original upper depositional surface of **QTa** has been substantially eroded, and its original depositional thickness is unknown. Pavement and soil development are younger than the latest erosional episode. Where **QTa** adjoins **Qao**, **QTa** is clearly at a higher grade and more eroded than **Qao**. **QTa** locally includes interbedded megabreccia deposits in the Blue Diamond area (Page and others, 1998) and in the west-central Willow Peak quadrangle. Megabreccia layers that have not been mapped separately crop out in the Mount Stirling quadrangle (N1/2 sec. 19, T. 18 S., R. 54 E.; sec. 1 and sec. 2, T. 18 S., R. 53 E.) and Horse Spring quadrangle (sec. 24, T. 18 S., R. 53 E.). Minimum exposed thickness of **QTa** is at least 50 m; base of unit is generally not exposed. In the Pahrump Valley, **QTa** includes gravel that contains sparse granitic clasts that were probably derived from the Kingston Range. Unit caps a chain of rounded hills that have a northwest alignment parallel to the trend of escarpments to the east. To the southeast along the general alignment of **QTa** within the Stump Spring quadrangle (McMackin, 1999), better exposures of this unit include abundant and lithologically varied volcanic clasts in beds striking northwest and dipping northeast beneath **Qb**. Malmberg (1967) reported a white to light-green tuff

that underlies these gravel deposits, but good exposures of the gravel or underlying stratigraphy were not noted in the map area. The flanks of the gravel-covered hills are draped with thick colluvium containing a younger eolian sand component. **QTa** in the subsurface is considered here to include basin-fill deposits that may be equivalent to parts of the Muddy Creek and Horse Spring Formations (Bohannon, 1984; Maldonado and Schmidt, 1991)

Footslope Colluvium and Alluvium

Qcf   **Hillslope colluvium and alluvium (Holocene and Pleistocene)**—Interbedded colluvial and debris-flow diamicton beds that grade into and are interbedded with alluvium of a wide age range. Unit generally occurs on lower, concave-upward footslopes (Peterson, 1981), which are typically partially dissected. Consists of angular and subangular gravel ranging in size from granules to boulders, generally supported by a matrix with variable proportions of sand, silt, and clay; matrix material inferred to be at least partly of eolian origin. **Qcf** is massive to finely bedded and locally includes multiple buried soils indicating Pleistocene episodes of pedogenesis and eolian deposition. Includes boulder levees that adjoin and are parallel to modern gullies and that were formed by debris flows. Includes deposits of possible glacial or periglacial origin in the southeast part of the Charleston Peak quadrangle (sec. 27, T. 19 S., R. 56 E.), just south and southwest and upslope of Big Falls. The contact with adjoining alluvium is well defined only where unit **Qcf** is truncated by younger alluvium. Exposed thickness ranges from 1 to 4 m

Fine-grained Deposits

Qd   **Dune sand (late Holocene)**—Noncemented fine sand in intermittently active to inactive dunes partially vegetated by mesquite. Most dunes in the Pahrump Valley and Corn Creek Springs area occur on and near northwest-trending escarpments. The dunes mapped by Bingler (1977), Matti and Bachuber (1985), and Matti and others (1987, 1993) in Las Vegas Valley have been mostly

obliterated by urbanization. Most dunes that still exist can be considered intermittently active, with evidence for recurrent activity during the late Holocene. In the Corn Creek Springs area, radiocarbon dates on archaeological hearths buried by dunes range from about 4,000 to 5,200 years B.P. (Williams and Orlins, 1963). In the Pahrump Valley, a TL date (LV-24, table 1) on dune material from about 3 m depth in a mesquite-covered dune that mantles a scarp is about 2 ka. Thickness ranges from 1 to about 5 m. Eolian sand is also mapped in this unit as sand ramps along fluvial scarps of Red Rock Wash. However, eolian sand is not mapped separately where it is a common component of surface and buried soils in alluvial gravels and in fine-grained deposits associated with past ground-water discharge

Qpy   **Modern playa sediment (late Holocene)—**Fine mud, silt, and clay of modern playa in Pahrump Valley, southwest corner of map area. Smooth, flat, high albedo surface has several parallel, straight lineaments apparent on aerial photographs that are provisionally interpreted as buried, probable right-lateral strike-slip faults of the State Line fault zone. Margins of playa are transitional to unit **Qfy** with greater amount of fluvial sand and fluvial morphology. Thickness unknown; only upper surface of unit is exposed

Qsyy   **Youngest spring deposits (late Holocene)—**Historic and prehistoric ground-water discharge areas, including spring mounds, composed of fine-grained calcareous to organic-rich silt, clay, and mud. Unit includes the area of historically active artesian springs that now includes the north (main) well field for the Las Vegas Valley Water District. Unit also includes some spring mound areas, including some that were historically active and some which yielded late Holocene dates younger than 2,000 years B.P. (Haynes, 1967; Bell and others, 1998, 1999)

Qfy   **Intermittently active fluvial fine-grained alluvium (late Holocene)—**Brown to gray sand, silt, mud, and interbedded gravel. Includes units F and G of Haynes (1967) and Quade (1986) and distal facies of alluvial fans. Erosional and depositional fluvial bar and channel morphology ranges from fresh to muted. **Qfy** includes a thin pediment veneer across older fine-grained deposits. This thin veneer of intermittently active fluvial sediment grades up-wash to coarser gravelly youngest

alluvium (**Qayy**), with which **Qfy** is partly correlative. Thickness ranges from less than 50 cm to greater than 150 cm; base is not exposed, especially down wash to southwest in the Pahrump Valley, where this unit grades into modern playa sediment (**Qpy**)

Qfo   **Older fine-grained deposits (Holocene)—**Non-cemented fluvial sand and mud. Erosional and depositional fluvial bar and channel morphology ranges from fresh to muted, especially relative to adjoining unit **Qfy**. Map unit in up-wash position may represent a thin pediment veneer across older fine-grained units. Thickness ranges from less than 50 cm to greater than 150 cm; base is not exposed

Qse   **Unit E—Young fine-grained deposits associated with past ground-water discharge (early Holocene to latest Pleistocene)—**Light-gray to light-brown unconsolidated silt, sandy silt, silty sand, and mud; locally light-greenish-gray mud. Locally contains abundant dark-gray peat, charcoal, and organic-rich horizons (black mats) (Quade and others, 1998), aquatic and terrestrial mollusc shells (Taylor, 1967; Quade, 1986; Quade and Pratt, 1989; Brennan and Quade, 1997; Bell and others, 1998, 1999), and fossils of late Pleistocene megafauna, including mammoth, bison, horse, and camel (Mawby, 1967). Numerous radiocarbon dates (Haynes, 1967; Quade, 1986; Quade and others, 1995, 1998; Brennan and Quade, 1997; Bell and others, 1998) range from about 8,000 to about 15,000 yrs BP, with a significant number of black mats yielding dates of about 10,000 to 11,000 yrs BP (Quade and others, 1998). We obtained thermoluminescence dates in this age range from a site (LV-14, LV-15, table 1) in which 5 radiocarbon dates in this age range were obtained on black mats by Bell and others (1998). Commonly exists as channel fills inset within unit D (as in Quade and others, 1998). Bedding defined by changes in color, organic content, grain size, and sedimentary structures. Ranges from massive to well bedded with fine-scale crossbeds. Unit E was originally defined and described by Haynes (1967). Includes non-fossiliferous eolian/phreatophyte flat facies of Quade and others (1995). Thickness is greater than 4 m; base commonly not exposed

Qsy   **Undivided young spring deposits (Holocene and late Pleistocene)—**Includes units **Qse** and **Qsyy**. Thickness is 1–4 m; base is locally not exposed

Qscd   **Unit C and D—Intermediate fine-grained deposits associated with past ground-water discharge (late Pleistocene)—**Top 1–2 m is characteristically resistant light-gray calcareous mud that is partially cemented with calcite which weathers to curving and branching to platy nodules; some of these may be trace fossils of cicada burrows (Quade, 1986). Grades downward to underlying unit C(?), which is tan-brown (more oxidized), about 2–3 m thick, fine sandy silt and mud. This middle brown unit may include one buried soil, marked by a slight increase in prismatic structure and darker tone. Includes occasional fluvial cross-bedding and common carbonate nodules. These units were first defined by Haynes (1967) in the northern Las Vegas Valley. In the Pahrump Valley, the intermediate brown unit overlies a sharp contact with a strongly developed buried soil that may correlate with unit A of Haynes (1967) in the Las Vegas Valley. Numerous radiocarbon dates on unit D range from about 16 to 30 ky B.P., with most around 25 ky B.P. At the type section of unit D defined by Haynes (1967) in trench K at the Tule Springs site, we obtained a TL date (LV-17, table 1) of 24–41 ka. Unit C is generally lacking in fossils, but Bell and others (1998) obtained a radiocarbon date on snails of 29 ky B.P.; at this site we obtained a TL date (LV-16, table 1) of 24–36 ka. At the type section of unit C defined by Haynes (1967) in trench K at the Tule Springs site, we obtained a TL date (LV-18, table 1) of 35–56 ka. TL dates on this unit in Pahrump Valley (LV-21, LV-22, table 1) are generally in this same time range. Observed thickness ranges from 2 to 6 m; base commonly not exposed

Qsu   **Undivided young and intermediate fine-grained deposits associated with past ground-water discharge (early Holocene and late Pleistocene)—**Includes units **Qse** and **Qscd**

Qsab   **Older fine-grained deposits units A and B (late? to middle Pleistocene)—**Unit B, locally 3–4 m thick, consists of fine-grained greenish-gray to brown muds and marls as defined by Haynes (1967) where well exposed at the Tule Springs archaeological site. Where exposed in nearly vertical exposures, this unit is not mappable at 1:100,000-scale. Locally fossiliferous, including molluscs and megafauna. We obtained a TL date (LV-20, table 1) of 89–144 ka on the type section of unit B at the Tule Springs site of Haynes (1967).

Unit B is underlain by unit A, which consists of a red-brown buried soil at least 1 m thick in silt and mud similar to unit B. The buried soil in unit A is clay-rich with a well-developed angular blocky structure and common manganese oxide stains on ped faces. We obtained a TL date (LV-19, table 1) of 131–225 ka on the type unit A at the Tule Springs site of Haynes (1967). Unit A may correlate in part with unit **Qso**. Unit **Qsab** is mapped along the Eglington fault scarp (from Haynes, 1967)

Qso   **Old fine-grained spring deposits (middle Pleistocene)—**Light-toned, variably cemented fine sand, mud, and marl, generally associated with areas of past ground-water discharge. Generally calcareous, with different degree of cementation between beds, ranging from weakly to strongly cemented. Subvertical tubules within densely cemented layers may be plant casts. In the south-central part of the Valley quadrangle (Lundstrom and others, 1998), where **Qso** is crossed by Interstate I-15, white calcareous layers are interbedded with reddish-brown, silty, gypsiferous beds about 30–60 cm thick with medium to strong blocky structure. These may represent buried soils. Two or three of these beds are exposed in eroded buttes south of Interstate I-15, where they are interbedded with strongly cemented carbonate-rich layers. At site LV-1 (SW ¼ of sec. 29, T. 19 S., R. 62 E.), a thermoluminescence date of 226 to 399 ka (table 1) was obtained on the uppermost reddish-brown bed at about 3 m below the carbonate-cemented top of the unit. **Qso** is geomorphically expressed as rounded, poorly exposed badlands and less commonly as resistant buttes mantled by thin colluvium, sheetwash, and eolian sand. Unit may partly correlate to unit Qx of Haynes (1967). Minimum exposed thickness is 5 m; base is generally not exposed

Basin fill of Browns Spring (Pleistocene and Pliocene?)—Fine-grained bedded mud, fine sand, marl, and limestone with minor interbedded gravel; mapped only in the Pahrump Valley. Upper, middle, and lower parts of unit were mapped based on position relative to a resistant limestone bed in the middle unit

Qby   **Upper part (middle Pleistocene)—**Fine-grained, brown to whitish-tan mud, marl, and fine sand interbedded with minor 10–80-cm-thick beds of pebble gravel. Gravel is similar to surficial gravel in that it is subangular to

subrounded and composed of limestone rock types probably derived from the Mesozoic and Paleozoic rocks of the Spring Mountains. Gravel probably represents the distal ends of alluvial fans interbedded with the fine-grained deposits. Fine-grained beds are variably cemented with calcite and include beds interpreted as pre-Wisconsinan ground-water discharge cycles (Spaulding and Quade, 1996). This part of the section is generally poorly exposed and covered by colluvium. At Browns Spring, in muds about 3 m above the top of the resistant **Qbw** bed, we obtained a TL date (LV-27, table 1) of 211–284 ka. Minimum exposed thickness is about 50 m, but this unit is disconformable with overlying units

Qbw   **Middle white limestone (middle Pleistocene)—** Massive, 2-4-m-thick densely cemented limestone that forms a resistant escarpment extending to northwest and southeast of Browns Spring in the southwest part of the map area. Included in this unit is calcite-cemented limestone pebble gravel 30–60 cm thick, which overlies the massive limestone bed. At Browns Spring the gravel appears to be enclosed in the top of the massive limestone. U-series analysis from this unit yielded a date of about 370 ka (MSQ4B, table 2). Thickness is about 5 m

QTbo   **Lower part (middle and early? Pleistocene to Pliocene?)—**Fine-grained brown to whitish-tan muds and marls interbedded with minor pebble gravel. Marls form resistant ledges in escarpment below **Qbw** and include plant casts (stems of unknown taxa), and probably represent intervals of ground-water discharge. **QTbo** also contains interbedded lenses and channels of pebble gravel composed of limestone rock types. We obtained TL dates of about 210–400 ka (LV-25, LV-26, table 1) from this unit. Exposed thickness at least 40 m; base of unit is not exposed

Qfw   **Fine-grained sediments of Whitney Mesa (middle Pleistocene)—**Light-toned, variably cemented, reddish-brown, fine sand, mud, and marl, with local interstratified gravel. Exposed along the prominent escarpment along the eastern margin of Whitney Mesa in the southeast part of the map area. Generally calcareous, but with highly variable degree of cementation of beds, ranging from weakly to strongly cemented, and locally very nodular. Much of the carbonate may be secondary and formed by evaporation of

ground water where it reaches the escarpment free face, as locally occurs in present conditions. Luminescence dating of a silty, fine sand near the top of the unit, about 2 m beneath the highly cemented caliche which overlies the unit, yielded a TL date of about 232–379 ka (LV-6, table 1), similar to U-series dates on a travertine vein in the caliche cap (WM1A-A, WM1A-B, table 2)

QTs   **Undivided fine-grained sediments of the Las Vegas Valley (Quaternary and Tertiary?)—**Light-toned, variably cemented and interstratified fine sand, mud, marl, and minor gravel, probably associated with areas of past ground-water discharge and with distal facies of alluvial fans. Generally whitish to light gray but includes interstratified light-greenish-gray to yellowish-gray to pinkish-gray fine sand and silt, reddish-orange fine sand, and red-brown mud. Proportion of interbedded gravel in subsurface increases toward the west (Donovan, 1996). Generally calcareous, with highly variable degree of cementation of beds, interbeds, and locally capping caliche, ranging from weakly to strongly cemented; locally gypsiferous. Subvertical tubules within densely cemented layers may be plant casts. Generally corresponds to QTs as mapped by Matti and Bachuber (1985) and Matti and others (1987, 1993), to QTss as mapped by Bingler (1977), and to Qx as mapped by Haynes (1967). In most of the mapped area, this unit is poorly exposed because of urbanization. Because of the poor exposure, lack of age control, local similarity to each of the previously described fine-grained units and to the Muddy Creek Formation (**Tm**), and existence near and potential juxtapositon by faults, this unit could represent a wide range of ages and correlate locally to any of the previously described fine-grained units or **Tm**. Thickness estimated (from geophysics; Langenheim and others, 1997) to be as much as 4–6 km in the Las Vegas and Pahrump Valleys

QTu   **Undifferentiated fan alluvium and basin-fill deposits (Quaternary to Miocene?)—**Shown only in cross sections. Includes alluvial fan gravels and interbedded sands of units **Qay** and **Qai** at surface, and probably units **Qao** and **QTa** at unknown depth

Basin-fill Deposits

Tm   **Muddy Creek Formation (Pliocene and upper Miocene)—**Reddish-brown to greenish-gray

gypsiferous mudstone and fine sandstone; found only in a few small exposures in the southeast part of the map area beneath pediment gravels flanking the southwest margin of Frenchman Mountain, but is more extensive and better exposed in the Lake Mead area (Bell and Smith, 1980; Bohannon, 1984). Castor and others (2000) estimated the Muddy Creek Formation to be as great as 300 m thick in the Frenchman Mountain quadrangle

Th   **Horse Spring Formation, undivided (Miocene and Oligocene)—**Adapted from Maldonado and Schmidt (1991) in the Gass Peak quadrangle. Exposed in Sheep and Las Vegas ranges where conglomerate and lacustrine members have been recognized.

Conglomerate member—Gray to brownish-gray, poorly sorted, massive to vaguely thick bedded, coarse conglomerate; moderately consolidated. Detritus derived from local Paleozoic and Late Proterozoic sedimentary rocks. Includes landslide megabreccia. Depositional top not exposed; much of upper part probably removed by erosion. Minimum exposed thickness is about 100 m, but member is probably thicker than 200 m.

Lacustrine member—White and subordinate variegated red, yellow, and brown, laminated to thick-bedded, tuffaceous claystone and siltstone representing shallow lake deposits; locally interbedded with sandstone and pebble-cobble conglomerate. Some lake beds are calcareous and gypsiferous; subordinate thin limestone beds have algal mat structure. Includes light-gray air-fall tuff beds less than 1 m thick. K-Ar age dates of interbedded volcanic units in the member in the Gass Peak basin range from 12 to 16 Ma (Guth and others, 1988). Base is not exposed: exposed thickness about 200 m

Thb   **Basalt unit of Horse Spring Formation (Miocene)—**Several thin flows of alkali basalt interbedded with the conglomerate member of the Horse Spring Formation; found only in the Las Vegas Range in the northeast corner of the map area (Gass Peak NE quadrangle). Consists of 25–30 percent olivine phenocrysts. Dated at 16.4±0.6 Ma (K-Ar whole-rock date; Feuerbach and others, 1993). Basalt flows are 10–20 m thick

Volcanic Rocks

Tmd   **Mount Davis Volcanics (Miocene)—**Mostly resistant, light- to dark-gray and tan, vesicular, flow-foliated, fresh, crystal-poor to crystal-rich dacite, andesite, and basaltic andesite lavas and flow breccia, and minor interbedded gray and reddish-brown, volcanic mudflow breccia, air-fall tuff, and fluvial pebbly sandstone and sandy conglomerate. Exposed only in the McCullough Range in southeastern part of the quadrangle, especially along the road to, and near, the radio towers in sec. 25, T. 22 S., R. 62 E. Upper parts of unit, downfaulted and exposed northeast of radio towers, include moderately resistant, light-gray and tan, dacitic ash-flow tuff containing 10–20 percent phenocrysts of plagioclase, sanidine, biotite, and hornblende; the tuff here is overlain by an aphanitic gray basaltic andesite flow. **Tmd** occupies broad stream channels that were cut in gently tilted and eroded rocks of the Patsy Mine Volcanics (**Tpm**). Unit was correlated by Bingler (1977) with the Mount Davis Volcanics of the Eldorado Mountains (Anderson, 1977), about 15 km southeast of the map area. In nearby areas to the south and east, mapped as informal andesite unit within the volcanic rocks of the McCullough Range (Anderson, 1977; Bell and Smith, 1980; and Boland, 1996). [40]Ar/[39]Ar dates of the unit, 40 km south of the Las Vegas quadrangle in the McCullough Pass area, range from 13.1 to 14.0 Ma (Faulds and others, 1999). Thickness is about 300 m

Tpm   **Patsy Mine Volcanics (Miocene)—**Moderately resistant, tan, red, gray, and yellow, amygdaloidal, mostly crystal-poor andesite and basaltic andesite lavas and flow breccia and interbedded tan and light-green volcanic mudflow breccia, fluvial conglomerate, and fluvial sandstone. Generally propylitically altered. Flows contain 5–20 percent phenocrysts of plagioclase, hornblende, pyroxene, and olivine. Exposed only in McCullough Range in southeastern part of the quadrangle. Correlated by Bingler (1977) with the Patsy Mine Volcanics of the Eldorado Mountains (Anderson, 1977), which are about 15 km southeast of the quadrangle. In nearby areas to the south and east, mapped as informal altered andesite unit within the volcanic rocks of the McCullough Range (Anderson, 1977; Bell and Smith, 1980; Boland, 1996). An [40]Ar/[39]Ar date of about 15.6 Ma was obtained from the unit 40 km south of the Las Vegas quadrangle in the McCullough Pass area (Faulds and others, 1999). Thickness is about 200 m

Mesozoic, Paleozoic, and Proterozoic Rocks

[Abbreviations for major Mesozoic thrust plates in the quadrangle: Wheeler Pass, WPT; Gass Peak, GPT; Valley, VT; Dry Lake, DLT; Lee Canyon/Macks Canyon, LMCT; Deer Creek, DCT; Keystone, KT; and Bird Spring, BST. For example, the Wheeler Pass thrust plate (WPT) occupies the hanging wall of the Wheeler Pass thrust fault.]

Ja **Aztec Sandstone (Jurassic)—**Exposed along the east flank of the Spring Mountains in the BST. Unit is tan, pale-red, and pink quartzose sandstone containing well-sorted, subrounded quartz grains, hematite and calcite cement, and tabular-planar crossbeds as great as 10 m thick. Top of unit is truncated by the Keystone thrust fault. The Aztec Sandstone is conformable with the underlying Jurassic and Triassic units. Maximum exposed thickness is about 650 m

JℝRkc **Kayenta Formation (Lower Jurassic), Moenave Formation (Lower Jurassic), and Chinle Formation (Upper Triassic), undivided—**Exposed along the east flank of the Spring Mountains (BST). Upper part includes about 180 m of crossbedded sandstone and siltstone equivalent to the Lower Jurassic Kayenta and Moenave Formations of the Colorado Plateau region. Chinle Formation (Upper Triassic) is about 280 m thick and consists of sandstone, siltstone, shale, and local conglomeratic lenses. Sandstone and siltstone is red and yellowish brown and is generally thin bedded. Shale is red, olive green, and brown. At base, includes 10–30-m-thick Shinarump Member, which consists of conglomerate, pebbly carbonate, and calcareous pebbly sandstone; conglomerate clasts are composed mainly of distinctive reddish-brown, light-brown, and yellowish-brown chert. Unit forms slope beneath the massive cliffs of the overlying Aztec Sandstone. Thickness of map unit is about 460 m

 Moenkopi Formation (Middle? and Lower Triassic)—Exposed in the Blue Diamond area in the low foothills of the easternmost Spring Mountains (BST) and in the eastern Spring Mountains (KT). Consists of upper red member and undivided lower unit. Combined thickness is about 585 m

ℝm **Moenkopi Formation, undivided (Middle? and Lower Triassic)—**Includes upper red member and lower unit combined, and is shown in cross section only (cross section *C–C'*, map sheet 1)

ℝmu **Upper red member (Middle? and Lower Triassic)—**This informal member consists of two parts. The uppermost part weathers light gray to grayish white and contains interbedded claystone, dolostone, limestone, siltstone, and shale. Lower part weathers mostly reddish brown and contains siltstone, sandstone, claystone, mudstone, and some thin beds of limestone and gypsum. Some siltstone beds contain ripple marks on bedding planes. About 335 m thick

ℝml **Lower unit (Lower Triassic)—**Includes from top to base, Virgin Limestone Member, lower red member, and Timpoweap Conglomerate Member. The Virgin Limestone Member contains thin-bedded, medium-gray to yellowish-brown fossiliferous limestone, sandy limestone, dolostone, calcareous sandstone, and siltstone. Limestone is commonly nodular due to burrowing, and contains some thin discontinuous chert layers. Fossils include brachiopods, gastropods, pelecypods, and pelmatozoan ossicles identified as *Pentacrinus* (Hewett, 1931). Conodonts in the Virgin Limestone Member are Early Triassic in age (Harris and others, in press). Lower red member consists of reddish-brown siltstone, sandstone, and shale, and thin beds of light-gray, yellowish-gray, and moderate-orange-pink dolomitic limestone, sandy limestone, and gypsiferous siltstone. Timpoweap Conglomerate Member is dominantly conglomerate with minor pebbly carbonate and calcareous sandstone beds. Conglomerate clasts are composed mostly of subrounded to subangular gray, pink, and pale-red limestone and dark-brown chert in a carbonate matrix, and beds are tabular and graded. Upper contact of Timpoweap is gradational with lower red member. At base, Timpoweap generally rests unconformably on Kaibab Formation (Lower Permian). In parts of the Mountain Springs, Lost Cabin Spring, and Griffith Peak quadrangles, however, the Timpoweap lies disconformably on Lower Permian redbed unit. Thickness of lower member is about 250 m

Pkt **Kaibab and Toroweap Formations, undivided (Lower Permian)—**Best exposures are in the Blue Diamond area (BST); also exposed in the eastern Spring Mountains (KT) and at Frenchman Mountain. The Kaibab Formation consists of the Harrisburg and Fossil Mountain Members, recognized in Blue Diamond area (Carr, 1992; Carr and McDonnell-Canan, 1992). Harrisburg

Member is about 75 m thick and consists of light-gray micritic limestone and dolostone and bedded gypsum (individual beds as great as 1 m thick); includes yellowish-gray-weathering limestone and dolomitic limestone and gypsiferous red shale and claystone; pectinid valves and pelmatozoan ossicles observed; chert common in nodules and layers. Fossil Mountain Member is about 75 m thick and consists of mostly medium-gray, thick-bedded fossiliferous limestone and dolomitic limestone with brown-weathering chert layers and nodules; includes brachiopods, bryozoans, and pelmatozoan columnals. Conodonts collected in the Blue Diamond area indicate the age of the Kaibab Formation is late Early Permian, as old as late Kungurian and as young as Roadian (Harris and others, in press). The Toroweap Formation consists of the Woods Ranch, Brady Canyon, and Seligman Members (Carr, 1992; Carr and McDonnell-Canan, 1992). Woods Ranch Member is about 30 m thick and consists of light-reddish-gray-weathering interbedded shale, dolostone, siltstone, and some gypsum. Brady Canyon Member is about 67 m thick and consists of medium-gray to pinkish-gray limestone and dolomitic limestone with common chert nodules and layers. Seligman Member is 60 m thick and consists of tan to red interbedded siltstone, sandstone, and gypsum. Some dolostone and dolomitic sandstone in upper part. Unit forms ledgy to massive cliffs. Combined thickness of the two formations is about 250–300 m in the Blue Diamond area, 0–170 m in the eastern Spring Mountains, and about 300 m at Frenchman Mountain

Pr   **Lower Permian redbeds—**Exposed in the Blue Diamond area along low hills that make up the easternmost Spring Mountains (BST), the eastern Spring Mountains (KT), and at Frenchman Mountain. Consists of red, white, and yellowish-orange crossbedded sandstone and some interbedded siltstone and shale; locally gypsiferous. Contact with overlying Toroweap Formation is sharp and disconformable. In Lovell Canyon area (KT), unit is reported to be gradational with underlying Bird Spring Formation (Ho, 1990). In the Blue Diamond area (Carr and others, 2000; McDonnell-Canan and others, 2000), unit was correlated with the Hermit Formation of the Grand Canyon and Colorado Plateau regions. Unit forms ledgy

slope. Unit about 300 m (Carr and others, 2000; McDonnell-Canan and others, 2000) to 580 m (Bohannon and Morris, 1983) thick in Blue Diamond area and about 400 m thick at Frenchman Mountain (Matti and others, 1993)

PMc   **Callville Limestone (Permian, Pennsylvanian, and Mississippian)—**Unit mapped only at Frenchman Mountain in the eastern part of the map area. Includes stratigraphic equivalents of the Pakoon Formation (Lower Permian) of McNair (1951) and the Indian Springs Formation (Upper Mississippian) of Webster and Lane (1967). Upper 270 m (Pakoon Formation equivalent) is generally medium- to light-gray, thin-bedded limestone, dolostone, and gypsum (Matti and others, 1993). Middle 220–297 m (Callville Limestone, Pennsylvanian) is thin- to thick-bedded, medium- to dark-gray, fossiliferous, finely to medium-crystalline limestone and dolostone (Langenheim and Webster, 1979; Matti and others, 1993). Also includes layers and nodules of dusky-yellowish-brown-weathering chert; brown-weathering sandstone beds common in upper half. The Upper Mississippian Indian Springs Formation is about 25–30 m thick at Frenchman Mountain; consists of thin-bedded sandstone, siltstone, limestone, dolostone, and intraclastic limestone conglomerate (Langenheim and Webster, 1979). Contact between Indian Springs Formation and underlying rocks of Yellowpine Limestone of Monte Cristo Group is sharp and disconformable. Maximum thickness of unit about 600 m (Matti and others, 1993)

Bird Spring Formation (Lower Permian to Upper Mississippian)—Bird Spring Formation subdivided into upper and lower informal members (see map units **Pbu** and **PMbl** below) in Las Vegas Range (VT and DLT) and in Lucky Strike and Lee Canyon areas in the northern Spring Mountains (DCT and LMCT). Elsewhere in the quadrangle, the Bird Spring Formation was mapped undivided (**PMb**). Bird Spring Formation map unit includes Indian Springs Formation of Webster and Lane (1967) at base

Pbu   **Upper member (Leonardian and uppermost Wolfcampian)—**Upper part of member mostly Leonardian in age and about 900 m thick; consists of cyclic thin- to thick-bedded limestone, silty limestone, and dolostone that contain layers and nodules of dusky-yellowish-brown chert as

well as beds of calcareous siltstone and brown-weathering sandstone. Fossils include brachiopods, bryozoans, corals, fusulinids, gastropods, and pelmatozoan fragments. Very latest Wolfcampian conodonts, including *Sweetognathus whitei*, are in base of upper part (Harris and others, in press), but are succeeded by Leonardian fusulinids such as *Schwagerina crassitextoria* in Lee Canyon area (Rich, 1961). Ho (1990) reported even younger Leonardian fusulinids, such as *Parafusulina apiculata*, in the Lee Canyon area. Lower part of member forms regional marker within the thick Bird Spring Formation, and it was described by Rich (1961, 1963) and Page (1992, 1993, 1998). Consists mainly of alternating thinly laminated beds of limestone, silty limestone, mudstone, calcareous siltstone, and chert. Limestone and mudstone are dark gray to olive gray; rocks are organic rich and locally pyritic. Silty limestone and calcareous siltstone laminae weather yellowish gray to moderate brown. Some beds are turbiditic and convolutedly bedded. Macrofossils are generally sparse and include brachiopods, echinoderms, ichthyoliths, and trilobites; microfossils are more common including locally abundant sponge spicules, less abundant conodonts, and poorly preserved radiolarians. Marker unit contains submarine debris-flow conglomerate in channels and sheets. Clasts in conglomerate are bioclastic and contain shallow-water forms such as corals, fusulinids, and pelmatozoan fragments. Phosphatic concretions 0.5–4 cm in diameter locally present at base of marker unit in Lucky Strike Canyon area and southernmost Las Vegas Range; concretions commonly cored by fish fragments. Marker unit thickens west-northwestward in southern Las Vegas Range, from 350 m to 600 m (Lundstrom and others, 1998). Rich (1963) reported the marker unit to be 666 m thick in Lee Canyon area. Total maximum thickness of upper member is about 1,500 m

PMbl   **Lower member (upper Wolfcampian to Chesterian)—**Limestone, dolostone, chert, siltstone, silty limestone, sandstone, and shale. Limestone is medium gray, light gray, and yellowish gray. Commonly arenaceous and bioclastic; some beds oolitic. Limestone finely to coarsely crystalline, thin to thick bedded; some beds contain planar laminations and trough crossbeds. Abundant

discontinuous layers and nodules of dark-gray, dusky-yellowish-brown-weathering chert; locally, some beds in member contain more than 50 percent chert. Member also contains medium-gray to yellowish-gray dolostone. Abundant fossils include brachiopods, bryozoans, corals (*Syringopora*, *Chaetetes*, and solitary rugosans), and pelmatozoan columnals; fusulinids in upper part. Member about 1,045 m thick in southern Las Vegas Range (Lundstrom and others, 1998) and 1,010 m in Lee Canyon area in the Spring Mountains (Rich, 1963)

PMb   **Bird Spring Formation, undivided (Lower Permian to Upper Mississippian)—**Consists of yellowish-gray, light-brown to gray, thin- to thick-bedded, bioclastic, and arenaceous limestone and dolostone. Also contains brown sandstone, gray to yellowish-gray calcareous siltstone, pale-red to light-gray shale, and layers and nodules of dusky-yellowish-brown chert; some beds have greater than 50 percent chert. Unit contains abundant macrofossils; fusulinids are found in much of the upper part of the Bird Spring Formation. Macrofossils include brachiopods, bryozoans, colonial and solitary corals, gastropods, pelecypods, pelmatozoan columnals, sponges, and trilobites. Unit forms ledgy slopes and, locally, massive cliffs. Maximum exposed thickness is 2,355 m in Lee Canyon area (LMCT) in the Spring Mountains (Ho, 1990; Rich, 1963); maximum exposed thickness in southern Las Vegas Range (VT) is about 2,500 m; total thickness in Lovell Canyon area (KT) is 1,300 m (Ho, 1990)

Indian Springs Formation (Chesterian)—Consists of interbedded bioclastic limestone, shale, and quartzite. Limestone is medium gray, grayish red, and moderate yellowish brown, mostly thin bedded, bioclastic, finely to coarsely crystalline. Shale is mostly in lower part of formation and is dusky red to grayish black. Quartzite is olive gray to light gray and composed of fine, subrounded, and moderately sorted quartz grains. Fossils include the biostratigraphically diagnostic Chesterian brachiopod *Rhipidomella nevadensis*, other spiriferid and productid brachiopods, bryozoans, solitary rugose corals, and pelmatozoan fragments. *Stigmaria* is in sandy and silty beds at base of formation in Gass Peak area (VT). Contact between Indian Springs and overlying Bird Spring Formations apparently gradational in

quadrangle; unit disconformably overlies Battleship Wash Formation or units of the Monte Cristo Group. Indian Springs Formation forms slope and is 52 m thick at type section at Indian Springs (WPT) (Webster, 1969), just north of quadrangle; about 60 m thick in southern Las Vegas Range (VT) (Lundstrom and others, 1998); 8-15 m thick in La Madre Mountain area (KT) (Axen, 1985)

Mm   **Monte Cristo Group (Upper and Lower Mississippian)**—Includes following formations, from top to bottom: Yellowpine Limestone, Bullion Limestone, Anchor Limestone, and Dawn Limestone; series and stage assignments based chiefly on conodont data (fig. 2; Harris and others, in press; Stevens and others, 1996). Map unit also includes Battleship Wash Formation of Webster (1969) overlying the Monte Cristo Group in southern Las Vegas Range (Lundstrom and others, 1998) and Lee Canyon area (LMCT) in the central Spring Mountains (Langenheim and Webster, 1979). Monte Cristo Group (exclusive of Battleship Wash Formation) about 304 m thick in northwestern Spring Mountains (WPT) (Vincelette, 1964), 300 m in La Madre Mountain area in the eastern Spring Mountains (KT) (Axen, 1985), about 253 m at Mountain Springs Pass (KT) (Langenheim and Webster, 1979), 400–480 m in southern Las Vegas Range (VT) (Lundstrom and others, 1998), and 242 m at Frenchman Mountain (Matti and others, 1993)

Battleship Wash Formation (lower Chesterian and upper Meramecian)—Exposed in southern Las Vegas Range (VT); a thin continuation of this unit was identified by Webster (1969) in Lee Canyon area in the Spring Mountains. Formation composed of limestone and minor quartzite. Limestone is medium dark gray to light olive gray, mostly coarsely crystalline, and thin to massive bedded. Basal part of formation consists of thin beds of quartzite and sandy limestone. Fossils in formation include solitary corals and spiriferid brachiopods. In southern Las Vegas Range, conodonts indicate that the formation is no older than late Meramecian and no younger than early Chesterian (Harris and others, in press). Upper contact usually covered, but where exposed, shale or quartzite of Indian Springs Formation disconformably(?) overlies bioclastic limestone of Battleship Wash Formation. Forms ledgy

cliffs. Formation 42 m thick in southern Las Vegas Range

Yellowpine Limestone (Meramecian and upper Osagean)—Medium-dark-gray to light-olive-gray, medium-crystalline, and thin- to thick-bedded limestone; contains sparse nodules of gray to brown chert. Formation distinct from other units of Monte Cristo Group in quadrangle because it contains unusually large solitary rugose corals (as much as 25 cm long), especially near its top. Also contains *Syringopora* and *Lithostrotionella* colonial corals, pelmatozoan columnals, and middle to late Osagean conodont *Eotaphrus burlingtonensis* at its base in southern Las Vegas Range (Harris and others, in press) and late Meramecian conodont *Cavusgnathus unicornis* near its top in Lee Canyon area in the Spring Mountains (Stevens and others, 1996). Only 12 m of unit exposed just north of quadrangle in Indian Springs area (WPT) (Stevens and others, 1996); unit 45 m thick at Mountain Springs Pass (KT) (Langenheim and Webster, 1979) and 77–80 m thick in southern Las Vegas Range (VT) (Lundstrom and others, 1998)

Bullion Limestone (lower Meramecian to middle Osagean)—Medium-light-gray encrinitic wackestone to grainstone typifies Bullion Limestone. Contains some beds of dusky-yellowish-brown-weathering chert. Fossils include abundant pelmatozoan ossicles and some brachiopods and solitary rugose corals. Conodonts indicate much of the formation is middle Osagean (*anchoralis-latus* Zone); Stevens and others (1996) reported *Eotaphrus burlingtonensis* and *bactrognathids* near base in Lee Canyon in the Spring Mountains, and Harris and others (in press) found *E. burlingtonensis* and *Hindeodella segaformis* s.f. (redeposited?) at top in southern Las Vegas Range. Contact between Bullion Limestone and Yellowpine Limestone gradational, and it is placed at base of limestone sequence that is darker gray, thinner bedded, and richer in solitary rugose corals than underlying beds of Bullion Limestone. Forms massive cliffs. Bullion Limestone is about 60 m thick in Indian Springs area (WPT) (Stevens and others, 1996), 118 m at Mountain Springs Pass (KT) (Langenheim and Webster, 1979), and 100–180 m in southern Las Vegas Range (VT) (Lundstrom and others, 1998)

Anchor Limestone (Osagean)—Alternating thin-bedded limestone and chert. Limestone is

medium gray to light olive gray, partly burrowed, and fine to medium crystalline. Chert is dark gray and weathers moderate orange to light brown. Anchor Limestone contains brachiopods, solitary and syringoporid corals, and pelmatozoan fragments. A 5–6-m-thick bluish-gray, bedded spiculitic chert marks base of Anchor in northwest Spring Mountains (WPT). Anchor Limestone forms ledgy cliffs and is 246 m thick in Indian Springs area (WPT), just north of Las Vegas quadrangle (Stevens and others, 1996), 67 m at Mountain Springs Pass, 80 m at Kyle Canyon (KT) (Langenheim and Webster, 1979), and 140-180 m in southern Las Vegas Range (VT) (Lundstrom and others, 1998)

Dawn Limestone (lower Osagean and Kinderhookian)—Dark-gray, brownish-gray, and light-olive-gray, thin-bedded to massive, bioclastic, oolitic limestone. Commonly contains elongate nodules and discontinuous layers of brown chert. Abundant macrofossils include brachiopods, bryozoans, solitary rugose corals, *Lithostrotionella* sp. corals, gastropods, and pelmatozoan ossicles. Contact with underlying Crystal Pass Limestone gradational through 2 m in Kyle Canyon area (KT). A disconformity separates the Dawn Limestone and the underlying Guilmette(?) Formation in the Las Vegas Range (VT), and a disconformity also separates the Dawn Limestone and the underlying Narrow Canyon Formation in northwest Spring Mountains (WPT) (fig. 2). Dawn Limestone is 75 m thick in Indian Springs area, just north of Las Vegas quadrangle (WPT) (Stevens and others, 1996), 24 to 50 m in Mountain Springs Pass and Kyle Canyon areas (KT) (Langenheim and Webster, 1979), and 70 m in southern Las Vegas Range (VT) (Lundstrom and others, 1998)

MDu **Mississippian and Devonian rocks, undivided (Lower Mississippian and Upper and Middle Devonian)**—In WPT and GPT, unit includes Lower Mississippian Narrow Canyon Formation and underlying Upper and Middle Devonian Guilmette Formation. As exposed in DCT and LMCT, unit includes Mississippian and Devonian Crystal Pass Limestone and underlying Upper and Middle Devonian Guilmette Formation. In eastern Spring Mountains (KT) and at Frenchman Mountain, includes Sultan Limestone

Narrow Canyon Formation (middle Kinderhookian)—Mostly burrowed, sandy

limestone and quartzite. Sandy limestone is laminated, medium gray to yellowish brown, and contains scattered planispiral gastropods, pelmatozoan ossicles, and subrounded intraclasts. Some poorly exposed laminated beds of pinkish-gray argillaceous shaly limestone are in upper part. Unit also includes brown quartzite beds that contain *Skolithus* burrows and laminated beds of sandy carbonate and dolomitic sandstone. Unit produces Kinderhookian conodonts as well as reworked middle and early Famennian conodonts (Harris and others, in press; F.G. Poole, written commun., 1997). Forms slope beneath cliff-forming Dawn Limestone; contact between these two units is sharp. Lower contact with Guilmette Formation is also sharp and disconformable. Unit 30–40 m thick in northwest Spring Mountains (WPT)

Crystal Pass Limestone (lower Kinderhookian and upper Famennian)—Light-gray, thin- to thick- and massive-bedded, planar- and locally cross-laminated micrite to mudstone containing sparse gastropods, intraclasts, and burrow mottling. At most locations in Las Vegas quadrangle, upper part of Crystal Pass contains the conodonts *Siphonodella praesulcata* and (or) *Si. sulcata* (fig. 2) indicating a very earliest or early Kinderhookian and (or) very latest Famennian age. Crystal Pass Limestone forms massive cliffs; it is about 70 m thick in Lucky Strike Canyon area (DCT) (Bereskin, 1982), 75 m in Kyle Canyon area (KT), 64 m at Mountain Springs Pass (KT) (Langenheim and Webster, 1979), and 60 m at Frenchman Mountain (Rowland and others, 1990)

Guilmette Formation (Upper and Middle Devonian)—Equivalent to Devils Gate Limestone mapped in northwestern Spring Mountains (WPT) and in part to Sultan Limestone mapped in central Spring Mountains (LMCT and DCT) (Burchfiel and others, 1974). Medium-dark-gray to light-brownish-gray dolostone and limestone and minor dolomitic quartzite. Limestone and dolostone beds are cyclic and commonly laminated and burrow-mottled. Fossils include brachiopods, solitary and colonial corals, gastropods, and *Amphipora* sp. as well as other stromatoporoids. Upper part of formation contains several brown dolomitic quartzite beds. Middle part of formation mostly cliff-forming, dark-gray dolostone with abundant stromatoporoids. Lower part of formation forms ledgy slope beneath

cliff-forming dolostone and contains interbeds of yellowish-gray and light-gray dolomicrite. Lower part is "yellow bed" of the Guilmette (Tschanz and Pampeyan, 1970) and marks the base of the Guilmette Formation in many parts of Clark and Lincoln Counties and elsewhere in Nevada. Guilmette Formation is 490 m thick in southern Sheep Range (GPT) (Maldonado and Schmidt, 1991), 365-450 m in northwestern Spring Mountains (WPT), and 380-420 m in central Spring Mountains (LMCT and DCT)

Sultan Limestone (Lower Mississippian to Middle Devonian)—In eastern Spring Mountains (KT) and at Frenchman Mountain, unit follows definition of Hewett (1931). Sultan Limestone mostly equivalent to but thinner than Guilmette Formation; includes, from top to base, Crystal Pass, Valentine, and Ironside Members. These members form a dark band across west face of Frenchman Mountain composed chiefly of dark-gray to brown dolostone with numerous stromatoporoids (Matti and others, 1993); these rocks lie unconformably on Nopah Formation (Upper Cambrian). Thickness of all three members at Frenchman Mountain is 190 m (Matti and others, 1993). In eastern Spring Mountains (KT), Crystal Pass Member consists of light-gray micritic limestone. Valentine and Ironside Members consist of cycles of thin-bedded, dark-gray limestone and thick-bedded, light-gray dolostone. Valentine Member, like upper part of Guilmette Formation, contains several brown dolomitic quartzite beds generally less than 1 m thick. Sultan is about 175-200 m thick in eastern Spring Mountains (Axen, 1985)

Dsi **Simonson Dolomite (Middle Devonian)**—Mapped only in northwest Spring Mountains (WPT) and southern Sheep Range (GPT). Upper part of Mountain Springs Formation (**DOm**) in eastern Spring Mountains (KT) and much of the Devonian rocks, undivided (**Du**), in central Spring Mountains (DCT and LMCT) include partial equivalents of the Simonson. Simonson also equivalent to Nevada Formation of Burchfiel and others (1974). Unit consists mostly of alternating beds of medium- to dark-gray and light-olive-gray, slightly argillaceous, finely crystalline, thin- to thick-bedded dolostone. Fossils mainly *Amphipora* sp. as well as bulbous and tabular stromatoporoids; the Givetian (late Middle Devonian) brachiopod

Stringocephalus sp. is near the top of formation in northwest Spring Mountains. Base of the formation consists of a 10-m-thick, brown dolomitic quartzite unit; quartz grains are fine to very fine, subrounded, and moderately well sorted. Forms ledgy cliffs. About 190 m thick in southern Sheep Range (Maldonado and Schmidt, 1991) and 120–130 m thick in northwest Spring Mountains (Vincelette, 1964)

Du **Devonian rocks, undivided (Middle and Lower Devonian)**—Unit mapped in central Spring Mountains (LMCT and DCT); includes strata below the basal yellow-bed sequence of Guilmette Formation (Upper and Middle Devonian) and above the Ely Springs Dolomite (Upper Ordovician). In LMCT, upper part is about 60 m thick and consists of alternating light- to dark-gray burrowed dolostone containing *Amphipora* sp., other stromatoporoids, and some brachiopods; probably equivalent with part of Simonson Dolomite. Lower part in LMCT is about 20 m thick and consists of light-gray-weathering, cryptalgal laminated and thin-bedded, micritic to finely saccharoidal dolostone containing discontinuous layers and nodules of chert; no macrofossils noted. These lower rocks are stratigraphically and lithologically similar to rocks that lie above the Silurian Laketown Dolomite and which we correlate with the Lower Devonian Sevy Dolomite in the northwest Spring Mountains (WPT) and southern Sheep Range (GPT). In Lucky Strike Canyon area (DCT), upper and lower parts are about 50 and 10 m thick, respectively. Map unit is 60 (DCT) to 80 m (LMCT) thick

DSu **Devonian and Silurian rocks undivided (Lower Devonian and Lower Silurian)**—Exposed in southern Sheep Range (GPT) and northwestern Spring Mountains (WPT). Unit includes Silurian Laketown Dolomite and overlying Lower Devonian strata. In Spring Mountains, upper 30 m probably include Lower Devonian strata likely equivalent to part of Sevy Dolomite; uppermost 17 m of these strata are light-brownish-gray-weathering, medium- to medium-dark-gray, laminated sandy dolostone containing anastomosing, medium-gray chert nodules and lenses. These rocks are equivalent to Lower Devonian cherty argillaceous unit, originally defined as an informal member of the Sevy Dolomite by Osmond (1962), but later redefined by Johnson and others

(1989) as a separate unit. Contact with overlying basal quartzite of Simonson Dolomite is sharp; as much as 3 m of relief observed locally along unconformity. Lower 13 m of Sevy Dolomite equivalent strata is light-gray, thin- to thick-bedded, finely saccharoidal dolostone containing cryptalgal laminations, fenestral fabric, and rare pelmatozoan ossicles. Dolostone is partly vuggy and burrow mottled, and contains scattered intraclasts. Beneath Lower Devonian strata, Laketown Dolomite displays its typical tri-part character that is widely recognized in eastern Great Basin: upper dark, middle light, and lower dark dolostone parts. Upper dark part is medium-dark-gray, mostly finely saccharoidal dolostone containing brachiopods, halysitid corals, and pelmatozoan ossicles. Middle part is light-gray to light-orange-gray, massive-bedded dolostone that contains silicified pentamerid brachiopods. Lower part is gray, vuggy and burrowed, finely saccharoidal dolostone with brachiopods, corals, pelmatozoan ossicles, and stromatoporoids; also contains some beds and nodules of dark-brown-weathering chert. The Laketown Dolomite is Early Silurian in age (at least late Llandoverian and early Wenlockian) based on conodonts collected in Corn Creek Springs quadrangle, southern Sheep Range (fig. 2, section 1). Forms ledge-and-slope topography. Map unit about 280 m thick in southern Sheep Range and 275 m in northwest Spring Mountains

DOm   Mountain Springs Formation (Middle Devonian, Upper Ordovician, and Lower Ordovician)—Exposed only in eastern Spring Mountains (KT). Defined by Gans (1974) and subsequently modified by Miller and Zilinsky (1981) based on conodont biostratigraphy. Hintzman (1983) and Cooper (1987), using Miller and Zilinsky's revisions, subdivided the Mountain Springs Formation into three informal members: (1) upper member of Middle and Early Devonian age; (2) middle member of Late Ordovician age; and (3) lower member of Middle and Early Ordovician age. New closely spaced conodont samples from Miller and Zilinsky's (1981) measured section, west-southwest of La Madre Spring, further refines age assignments. Near La Madre Spring (fig. 2, section 6), Mountain Springs Formation about 295 m thick. Upper

member, about 60 m thick, is restricted to Middle Devonian as only early Middle Devonian (Eifelian) conodonts were found (conodonts identified by Miller and Zilinsky, 1981, as Early Devonian in age are Middle Devonian forms), and it is equivalent to part of Simonson Dolomite. Upper member characterized by cycles of dark- to medium-light-gray and brownish-gray, burrow-mottled, finely to coarsely saccharoidal dolostone grading up into light- to very light gray and pinkish-gray, thin- to thick-bedded, laminated, finely saccharoidal dolostone to dolomudstone with scattered rip-up clasts. Lower part of cycle locally contains disarticulated, thick-shelled pentamerid brachiopods and stromatoporoids and, more rarely, bryozoan, coral, and pelmatozoan fragments. Middle member of Late Ordovician age, 6–8 m thick, disconformably overlies and underlies lower and upper members, and is equivalent to part of Ely Springs Dolomite. Beds are poorly exposed, light-pinkish-gray-weathering, brownish-gray, medium-bedded, burrow-mottled to even-bedded, bioclastic (mainly pelmatozoans), moderately saccharoidal dolopackstone. Lower member, about 225 m thick, contains abundant and diverse Early Ordovician conodont assemblages in upper 175 m (this report, fig. 2, section 6; Harris and others, in press) but lower 50 m paleontologically unproductive and may include rocks of Late Cambrian age; thus lower member equivalent to part of Pogonip Group. Although considerably thinner than the Lower Ordovician part of Pogonip Group in southern Sheep Range (GPT), many of the low-latitude, shallow-water Laurentian province (North American Midcontinent conodont province of older usage) Lower Ordovician conodont zones are recognized in the La Madre Spring section (fig. 2, section 6). Lower member consists of light-gray to pinkish-gray, platy to massive-bedded, mostly burrowed, finely to coarsely saccharoidal, dolomitized wackestone, packstone, and mudstone with common orange-weathering argillaceous partings and some chert. Rocks contain stromatolites, cryptalgal lamination, intraclasts, and fenestral fabric common in middle and lower part of member; locally contains oncoids and intraclasts. Notable karstification features including solution surfaces, collapse breccia beds (prominent collapse breccia containing associated sinkhole-fill deposits that include

rounded chert clasts averaging 2 cm at 168–174 m below top of formation). Macrofossils include large gastropods (averaging 2 cm in diameter) and sponges. Thickness and age limits of informal members will likely vary from place to place in KT because older and (or) younger beds may have been differentially removed or preserved along the major unconformities that separate these members

Oes   Ely Springs Dolomite (Upper Ordovician)—Upper 20–25 m is generally light-olive-gray, thin- to thick-bedded, burrow-mottled, variably bioclastic, finely saccharoidal dolostone containing ooids, oncoids, and pelmatozoan ossicles. Most of Ely Springs Dolomite is medium-dark-gray, burrow-mottled, irregularly thin- to thick-bedded with common planar laminations, finely saccharoidal dolostone; fossils include brachiopods, colonial (*Halysites* sp. and favositids) and solitary corals, gastropods, pelmatozoan ossicles, and stromatoporoids (including amphiporids). Lower contact with Eureka Quartzite is disconformity; lower Middle Ordovician quartzite of Eureka is overlain by Upper Ordovician dolostone of Ely Springs Dolomite. Formation is 150 m thick in southern Sheep Range (GPT), 120 m in northwest Spring Mountains (WPT) (Vincellete, 1964), 76–107 m in LMCT (Vincellete, 1964), and 90 m thick in DCT

Oe   Eureka Quartzite (Middle Ordovician)—Light- to moderate-brown-weathering, white to light-brown, thin- to thick-bedded, fine- to medium-grained, rounded to subrounded, moderately well sorted quartzite and friable sandstone. Contains tabular planar crossbeds and, less commonly, trough crossbeds; crossbed sets average about 0.4 m thick; also contains *Skolithus* burrows. Minor sandy carbonate beds. Local collophane nodules in Lee Canyon and Lucky Strike Canyon areas (Ross, 1964). Forms rounded cliffs. Unit 6 m thick in Lucky Strike Canyon area (DCT) and as much as 120 m thick in both northwestern Spring Mountains (WPT) (Vincellete, 1964) and southern Sheep Range (GPT) (Maldonado and Schmidt, 1991). Unit is absent in the eastern Spring Mountains (KT) and Frenchman Mountain

OЄp   Pogonip Group (Middle Ordovician to Upper Cambrian)—Includes Antelope Valley and Goodwin Limestones. Pogonip Group is 710 m thick in southern Sheep Range (GPT),

about 700 m in northwestern Spring Mountains (WPT) (Vincelette, 1964), and about 360–400 m in central Spring Mountains (LMCT and DCT)

Antelope Valley Limestone (Middle and Lower Ordovician)—Dark-yellowish-orange- to dark-yellowish-brown, medium-dark- to medium-gray, massive- to medium-bedded, coarse- to fine-grained, bioclastic limestone with locally abundant ooids, oncoids, and intraclasts; beds are commonly burrow mottled to churn burrowed. Also contains dark-yellow-orange silty to argillaceous laminae and nodules and layers of brown chert. Fossils include brachiopods, orthocone cephalopods, gastropods (*Palliseria* sp. and *Maclurites* sp.), pelecypods, pelmatozoan columnals, sponges (*Receptaculites* sp.), and trilobites. Rocks in the upper part of formation contain shallowing-upward cycles of bioclastic-oolitic packstone/grainstone grading up into bioturbated lime mudstone, laminated lime mudstone, and, finally, mud-cracked, cryptalgal laminated, argillaceous dolomitic lime mudstone. Ooids, oncoids, and intraclast beds (shoal facies) typify lower 370 m of Antelope Valley in southern Sheep Range. Forms cliffs

Goodwin Limestone (Lower Ordovician and Upper Cambrian)—Medium-gray to grayish-orange-weathering, thick-bedded, burrow-mottled to laminated, bioclastic (chiefly trilobites and brachiopods), mostly medium grained grainstone and packstone that characteristically contain brown-weathering chert layers and nodules, moderate-reddish-orange silty partings, and intraclastic beds; commonly oolitic. Samples of the Goodwin Limestone from several localities in the Las Vegas quadrangle have produced very late, but not latest Cambrian conodonts from its lower beds (fig. 2, sections 1 and 2; Harris and others, in press). Contact with underlying Nopah Formation is sharp

Єn   Nopah Formation (Upper Cambrian)—Dolostone, dark- to light-gray, mottled, medium-crystalline and thin- to thick-bedded with discontinuous layers and nodules of dusky-yellowish-brown-weathering chert. Alternating light- and dark-gray beds form distinctive color bands, similar to those in Bonanza King Formation, but bands are thicker and more prominent. Stromatolites, oncoids, and brachiopod fragments are common, pelmatozoan ossicles less common. Dunderberg Shale

Member makes up the base of unit and consists of silty limestone and olive-green shale; common trilobite fragments (*Dunderbergia* sp.) and grayish-orange intraclasts. Dunderberg Shale Member is conformable with underlying Bonanza King Formation. Most of the Nopah Formation forms massive cliffs, but Dunderberg Shale Member forms ledgy slopes. Map unit is 380 m thick in southern Sheep Range (GPT) (Maldonado and Schmidt, 1991), about 360 m in the northern Spring Mountains (WPT) (Vincelette, 1964), 345 m in eastern Spring Mountains (KT) (Axen, 1985), and 33 m at Frenchman Mountain (Matti and others, 1993)

€bk   **Bonanza King Formation (Upper and Middle Cambrian)—**Consists of the Banded Mountain and Papoose Lake Members. The Banded Mountain Member generally is light- to dark-gray, fine- to medium-crystalline dolostone and limestone; alternating light- to dark-gray colors give the member its distinctive banded appearance. Dark-brown to orange burrow mottling is prevalent in the member, and several intervals have chert nodules and discontinuous layers. Base of the member is marked by the regionally extensive "silty unit" (Barnes and Palmer, 1961; Guth, 1980; Gans, 1974), which is red-, orange-, and tan-weathering silty dolostone about 15 m thick, and contains Middle Cambrian trilobites (Guth, 1980). The Papoose Lake Member is mostly dark-gray dolostone with distinctive brownish-gray to grayish-orange burrow mottles; also includes sparse limestone and silty dolostone beds. Map unit is 900 m thick in the southern Sheep Range (GPT) (Maldonado and Schmidt, 1991), 853 m in the northwestern Spring Mountains (WPT) (Vincellette, 1964), about 445 m in the eastern Spring Mountains (KT) (Axen, 1985), and 484 m at Frenchman Mountain (Matti and others, 1993)

€c   **Carrara Formation (Middle and Lower Cambrian)—**Interbedded limestone, siltstone, sandstone, and shale. Upper part is gray, platy limestone containing yellow- and red-weathering silt or clay partings. Contact with overlying Bonanza King Formation is marked by a change from slope-forming, yellowish-weathering silty limestone of the Carrara Formation to more massive, cliff-forming, gray carbonate rocks of the Bonanza King Formation. Unit is 457 m

thick in northwest Spring Mountains (WPT) (Vincelette, 1964), and about 200 m of unit is exposed in southern Las Vegas Range (GPT) (Maldonado and Schmidt, 1991)

€b   **Bright Angel Shale (Middle and Lower Cambrian)—**Exposed only at Frenchman Mountain. Includes rocks equivalent to parts of the Chisholm Shale, Lyndon Limestone, and Pioche Shale. Chisholm Shale equivalent consists of 25 m of greenish-gray shale and minor limestone and limy shale; *Glossopleura* sp. trilobites reported by Palmer (1981). Lyndon Limestone equivalent consists of 30 m of dark-gray, thin-bedded limestone containing light-gray to brownish-gray burrow mottling. Pioche Shale equivalent about 128 m thick and composed mostly of olive-green phyllitic shale with local thin beds of dark-reddish-brown to purple sandstone; contains olenellid trilobite fragments (Pack and Gayle, 1971)

€t   **Tapeats Sandstone (Lower Cambrian)—**Exposed only at Frenchman Mountain. White to light-brown sandstone and quartzite with conglomeratic beds at base; thin- to thick-bedded, with trough cross bedding. Thickness is about 48 m (Hardy, 1986)

€Zw   **Wood Canyon Formation (Lower Cambrian and Late Proterozoic)—**Exposed in northwest Spring Mountains (WPT) and southern Las Vegas Range (GPT). Quartzite, sandstone, siltstone, and shale. The quartzite is similar to that in the underlying Stirling Quartzite, but it forms slopes in contrast to the cliff-forming Stirling Quartzite. Top of unit consists of a 7–8-m-thick ledge of white vitreous quartzite correlative with the Zabriske Quartzite (Vincellette, 1964). Shaly and silty units increase in abundance in upper part of formation. Basal part of formation is tan-weathering silty sandstone that forms a slope above the resistant Stirling Quartzite. Olenellid trilobites reported from near top of formation in Sheep Range, north of the quadrangle (Guth, 1980). Thickness from 640 to 820 m in Spring Mountains (Vincelette, 1964); maximum exposed thickness in Las Vegas Range is 450 m (Maldonado and Schmidt, 1991)

Zs   **Stirling Quartzite (Late Proterozoic)—**Exposed in northwest Spring Mountains (WPT) and in southern Las Vegas Range (GPT). Unit is mostly purple, pink, maroon, gray, and white conglomeratic quartzite, quartzite, and sandstone, with minor beds of light- to dark-brown micaceous siltstone. Lowermost

20 m mostly white quartzite. Unit locally metamorphosed to sub-greenschist and lower greenschist facies in parts of northwestern Spring Mountains. Unit forms cliff. Thickness in Spring Mountains about 975 m (Nolan, 1924). Only the uppermost 100 m is exposed in Las Vegas Range (Maldonado and Schmidt, 1991)

Zj   **Johnnie Formation (Late Proterozoic)—**Exposed only in northwest Spring Mountains (WPT). Only the uppermost 600 m of unit is exposed in quadrangle. Quartzite, sandstone, siltstone, shaly siltstone, and shale. Uppermost 60 m is micaceous siltstone with interbeds of dark-brown- to pale-red-weathering calcareous sandstone. Sandstone beds increase in number upward to become transitional with the overlying Stirling Quartzite. Next lower 30 m is green-weathering phyllitic shale and siltstone with interbeds of brown sandstone. Below, the next lower 90–120 m is dark- to light-brown-weathering dolomitic sandstone, siltstone, and quartzite. Underlying this unit is a marker horizon that consists of a 6-m-thick ledge of light-tan-weathering oolitic dolostone underlain by a 15-m-thick slope-forming blue-green shaly siltstone. Lowest exposed beds are reddish-brown, tan, gray, and green quartzite, sandstone, and siltstone. Formation is reported to be transitional with overlying Stirling Quartzite (Vincellette, 1964); base is not exposed

Xg   **Gneiss (Early Proterozoic)—**Exposed only at Frenchman Mountain. Gray microcline-quartz-biotite garnet gneiss and banded pink to white leucocratic gneiss. Gneiss is cut by coarse-grained biotite granite. Rowland (1987) correlated unit with Vishnu Schist in Grand Canyon area, with an age of 1.7 Ga in the Grand Canyon area (Elston, 1989)

Stratigraphy of the Las Vegas 30′ × 60′ Quadrangle

[Abbreviations for major Mesozoic thrust plates in the quadrangle: Wheeler Pass, WPT; Gass Peak, GPT; Valley, VT; Dry Lake, DLT; Lee Canyon/Macks Canyon, LMCT; Deer Creek, DCT; Keystone, KT; and Bird Spring, BST. For example, the Wheeler Pass thrust plate (WPT) occupies the hanging wall of the Wheeler Pass thrust fault.]

Proterozoic Rocks

Proterozoic rocks in the quadrangle are exposed at Frenchman Mountain and in the northwest Spring Mountains (WPT) and the southern Las Vegas Range (GPT). The oldest rocks in the quadrangle are gneissic crystalline rocks of Early Proterozoic age exposed at the base of Frenchman Mountain. Although these rocks have not been dated, similar rocks in the Grand Canyon area are 1.7 Ga (Elston, 1989). At French-man Mountain, the Early Proterozoic crystalline rocks are unconformably overlain by the Cambrian Tapeats Sandstone; this impressive unconformity represents a geologic time gap of about 1 b.y. Geophysical modeling suggests similar Early Proterozoic rocks may underlie the Spring Mountains. Such modeling is based on isostatic gravity and magnetic anomalies centered over the range (Langenheim and others, chapter B of this report).

Late Proterozoic rocks are exposed only in the northwest Spring Mountains (WPT) and southern Sheep Range (GPT) and include the Johnnie Formation, Stirling Quartzite, and the lower part of the Wood Canyon Formation. These are mostly clastic rocks deposited in shallow marine waters along a pas-sive continental margin in what is now western North America (Stewart, 1976; Stewart and Poole, 1972). Late Proterozoic rocks are interpreted to represent initial deposits of the Cordil-leran miogeocline (Stewart, 1972, 1976; Stewart and Poole, 1972), a broad paleogeographic feature that extends from Canada to northern Mexico. The miogeocline is characterized by a westward-thickening sequence of Late Proterozoic clastic and lower Paleozoic carbonate shelf rocks deposited along the North American continental margin as a result of Late Protero-zoic rifting (Stewart, 1976).

Paleozoic Rocks

Paleozoic rocks make up the greatest proportion of bedrock units in the quadrangle. Their distribution is important because they are part of the Nevada carbonate rock province (Dettinger and others, 1995) and are the major ground-water aquifers in eastern Nevada. Cambrian through Devonian rocks form the upper part of the miogeoclinal sequence. Meso-zoic thrust faults in the quadrangle juxtapose miogeoclinal sequences above progressively shallower shelf sequences, from west to east. For example, rocks of the WPT and GPT are miogeoclinal sequences that have been thrust above inner miogeoclinal sequences in the LMCT, VT, and DCT. Inner miogeoclinal rocks were thrust above cratonic margin rocks in the KT. At Frenchman Mountain, these rocks represent a cratonic sequence (Rowland, 1987).

Burchfiel and others (1974), Cooper (1987), and Cooper and Edwards (1991) discussed major unconformities within the Cambrian through Devonian sequence in the Spring Moun-tains; these unconformities account for dramatic stratigraphic thinning of these units across the Las Vegas quadrangle. We further refined the stratigraphic relations along these uncon-formities based on conodont biostratigraphy; these results

are shown in figure 2. In the WPT and GPT (fig. 2, sections 1 and 2), unconformities exist between Middle Devonian and underlying Lower Devonian units, Silurian and underlying Upper Ordovician units, and between Upper Ordovician and Middle Ordovician rocks. At least two of these unconformities are traced eastward to the cratonic margin sequence in the KT (fig. 2, sections 5 and 6), where they exist within the Ordovician to Devonian Mountain Springs Formation. Here, Middle Devonian Simonson Dolomite equivalent rocks overlie a very thin Upper Ordovician section correlative with the Ely Springs Dolomite, and these rocks, in turn, overlie Lower Ordovician rocks equivalent to the lower part of the Pogonip Group. Cooper and Edwards (1991) described the lower unconformity between Pogonip Group and Ely Springs Dolomite equivalent rocks as a type-1 sequence boundary marking the top of the Sauk sequence of Sloss (1963), and they described the upper unconformity between Upper Ordovician and Middle Devonian rocks as a type-1 sequence boundary marking the top of the Tippecanoe sequence of Sloss (1963).

Cambrian rocks in the quadrangle include part of the Wood Canyon, Carrara, Bonanza King, and Nopah Formations. Also included are the Tapeats Sandstone and Bright Angel Shale, which represent the cratonic sequence at Frenchman Mountain. Stewart (1970) correlated the Tapeats Sandstone with part of the Wood Canyon Formation, and the Bright Angel Shale with part of the Carrara Formation. These rocks are mostly clastic units consisting of sandstone, quartzite, conglomerate, siltstone, and shale, but they also include subordinate carbonate rocks. During Middle Cambrian time, a gradual change occurred from clastic to predominantly carbonate deposition along the miogeocline. Middle and Upper Cambrian rocks of the Bonanza King and Nopah Formations consist of dolostone, limestone, silty limestone and dolostone, shale, and siltstone, and these formations formed mostly in inner-shelf depositional environments that fluctuated from near-shore intertidal, shoal, and lagoonal environments to off-shore inner shelf settings. These rocks typically contain stromatolites and trilobites. The Dunderberg Shale Member of the Nopah Formation is a distinctive marker unit in the sequence and consists of green to brown shale and siltstone and limestone, with abundant *Dunderbergia* sp. trilobite fragments.

Ordovician rocks in the Las Vegas quadrangle include the Pogonip Group, Eureka Quartzite, and Ely Springs Dolomite. Miogeoclinal sections of the Pogonip Group are in the northwest Spring Mountains (WPT) and in the southern Sheep Range (GPT), where the group is about 700 m thick. Pogonip Group equivalent rocks thin to about 225 m in the eastern Spring Mountains (KT) in the cratonic margin sequence, represented by the lower member of the Mountain Springs Formation; Ordovician rocks are absent at Frenchman Mountain.

Although traditionally assigned an Ordovician age, samples of the Goodwin Limestone of the Pogonip Group in the quadrangle have produced very late, but not latest Cambrian conodonts from its lower beds (fig. 2, sections 1 and 2). The Antelope Valley Limestone of the Pogonip Group contains abundant, generally diverse conodont assemblages except in

rocks deposited in restricted very shallow water facies (such as cryptalgal laminated, mudcracked, lime mudstones). Nearly all of the low-latitude, shallow-water Laurentian province conodont zones of the Lower to lower Middle Ordovician (*R. manitouensis* Zone to at least *H. holodentata* Zone) have been identified, thus allowing precise correlation across thrust plates that represent significant variations in depositional environment (fig. 2). Conodont samples from the Pogonip Group typically represent wide-ranging open marine depositional environments ranging from deep-water open ocean to shallow inner-shelf settings. Distinctive fossils in the Antelope Valley Limestone are the spongiomorph *Receptaculites* sp. and the gastropods *Palliseria* sp. and *Maclurites* sp. *Palliseria* sp. are abundant, large (as much as 8 cm across), and magnificently exposed on the south side of Yucca Gap in the southern Sheep Range (map A).

Rocks of the Eureka Quartzite conformably overlie Pogonip Group strata in the quadrangle. The Eureka Quartzite is a superb Paleozoic marker unit in the Great Basin. The light-tan quartzite contrasts sharply with the dark-gray dolostone of the overlying Ely Springs Dolomite. In most of the quadrangle, the unit consists of fine- to medium-grained orthoquartzite, but it also includes some sandstone and local sandy carbonate beds. In the western part of the quadrangle (WPT and GPT), the Eureka attains a maximum thickness of about 120 m. The Eureka Quartzite is 80 m thick in the LMCT but thins to only 6 m in the Lucky Strike Canyon area, where it is in the upper plate of the Deer Creek thrust fault. In the Apex area of the Dry Lake Range (in the adjacent Lake Mead 1:100,000-scale quadrangle), the Eureka Quartzite is also thin, only 11 m, where it is in the upper plate of the Dry Lake thrust fault. In addition to similar thickness, Ross (1964) identified collophane nodules in the Eureka Quartzite in both the Lucky Strike Canyon and Apex sections. Based on these similarities, Ross (1964) suggested these two sections palinspastically restore along the Las Vegas Valley shear zone. Ross' interpretation supports correlation of the Deer Creek thrust with the Dry Lake thrust as proposed in this report and in Lundstrom and others (1998). The Eureka Quartzite is absent in the eastern Spring Mountains (KT) and at Frenchman Mountain.

The Eureka Quartzite is disconformably overlain by the Ely Springs Dolomite of Late Ordovician age. These rocks are mostly dark-gray, bioclastic dolostone, and yield conodonts no older than middle Edenian and no younger than middle Maysvillian at the base and could be as young as Richmondian at the top of the formation (Harris and others, in press). Because of the regional extent and relatively rapid transgression represented by the unit, the base of the Ely Springs Dolomite was chosen as the primary datum, or tie line, for correlating Paleozoic units across major Mesozoic thrust plates in the quadrangle (fig. 2).

Silurian rocks in the quadrangle appear to be limited to the northern Spring Mountains (WPT) and southern Sheep Range (GPT). The Laketown Dolomite displays its typical tri-part character that is widely recognized throughout the Basin and Range Province; upper dark, middle light, and

lower dark dolostone. Here, as elsewhere in the region, the Laketown Dolomite has halysitid corals and silicified pentemarid brachiopods. The formation is about 250 m thick in the quadrangle and is Early Silurian in age, at least late Llandoverian and early Wentlockian on the basis of conodonts collected in the southern Sheep Range (fig. 2, section 1; Harris and others, in press). Conodonts indicate that rocks of the Laketown Dolomite formed in a shallow-water marine-shelf depositional setting. The Laketown Dolomite disconformably underlies probable Lower Devonian strata equivalent to the Sevy Dolomite, and disconformably overlies Upper Ordovician strata of the Ely Springs Dolomite. The Laketown Dolomite is presumably absent in thrust plates east of the WPT and GPT, where it was removed along a regional pre-Devonian unconformity.

Devonian rocks are widespread in the quadrangle and are mostly limestone and dolostone deposited on a shallow-water carbonate shelf that fluctuated between normal- and restricted-marine depositional settings. Dolomitic quartzite units are present at several horizons in the Devonian sequence, at the base of the Middle Devonian Simonson Dolomite and in the upper part of the Guilmette Formation and Sultan Limestone. Stromatoporoids are notable in dolostone units of the Simonson Dolomite and Guilmette Formation and Sultan Limestone. Devonian units in the quadrangle have a maximum thickness of 650 m in the southern Sheep Range and a minimum thickness of about 200 m at Frenchman Mountain.

Lower Devonian rocks are exposed in the northwest Spring Mountains (WPT) and southern Sheep Range (GPT) and in the central Spring Mountains (LMCT and DCT). Based on the presence of algal laminations and fenestral fabric and on the general absence of conodonts, these rocks were deposited in a shallow-water, partly restricted carbonate shelf environment. In the WPT and GPT, the sequence includes rocks equivalent to the Sevy Dolomite and cherty argillaceous unit of Johnson and others (1989). These rocks have a maximum thickness of 50 m, and the top of the sequence is truncated along an erosional surface with local relief of about 3-5 m. Above this unconformity is a distinctive quartzite unit that marks the base of the Middle Devonian Simonson Dolomite in the quadrangle, as well as in much of eastern Nevada. In Lee and Lucky Strike Canyons (LCMT and DCT), the cherty argillaceous unit is absent, and Lower Devonian rocks are only 10–20 m thick, respectively. In the eastern Spring Mountains (KT) and at Frenchman Mountain, Lower Devonian rocks are absent.

Middle and Upper Devonian units in the quadrangle include the Simonson Dolomite, the Guilmette Formation, and the lower part of the Sultan Limestone. The Middle Devonian Simonson Dolomite is exposed in the WPT and GPT; equivalent rocks are present in thrust sheets to the east (LMCT, DCT), where the basal quartzite unit is absent and correlative strata are much thinner and more typical of an inner miogeoclinal facies. Simonson Dolomite equivalent rocks are only 60 m thick in the KT and are represented by the upper member of the Mountain Springs Formation that formed in a cratonic

margin setting. The Simonson Dolomite is Eifelian and early Givetian in age in most parts of the quadrangle (Harris and others, in press). *Stringocephalus* brachiopods (Givetian) are found at the top of the formation in the northwest Spring Mountains (WPT).

The Middle and Upper Devonian Guilmette Formation and the lower part of the Sultan Limestone are, for the most part, correlative lithologically. In this report, the Guilmette Formation consists of rocks of miogeoclinal to inner miogeoclinal facies in the WPT-GPT and LMCT-DCT, respectively, and the Sultan Limestone consists of rocks of cratonic margin facies in the KT and cratonic facies at Frenchman Mountain. The lower part of the Sultan Limestone is thinner than the Guilmette Formation and lacks the yellow bed unit (Tschanz and Pampeyan, 1970) that defines the lower part of the Guilmette Formation throughout the eastern Basin and Range province.

In most parts of the quadrangle, the Crystal Pass Limestone overlies the Guilmette Formation, but in the KT and at Frenchman Mountain, it is the upper member of the Sultan Limestone as defined by Hewett (1931). The Crystal Pass Limestone forms a distinct mappable unit, and its light-gray appearance and overall micritic texture make it a good marker bed; for these reasons, the unit deserves formation rank. Although originally defined as Devonian in age (Hewett, 1931), conodonts establish that it is also partly early Kinderhookian in age in southeastern Nevada (Harris and others, in press; Stevens and others, 1996). A disconformity separates the upper Famennian lower part of the Crystal Pass from the lower Kinderhookian upper part of the formation. In the southern Las Vegas Range (VT), rocks that are below the Dawn Limestone of the Monte Cristo Group and above the Guilmette Formation are atypical of the Crystal Pass Limestone and contain quartzite beds, sandy ooid limestone, and burrowed dolostone. These rocks were deposited in a very high energy depositional environment in contrast to the laminated micrite that typifies the Crystal Pass Limestone elsewhere in the Las Vegas quadrangle. These stratigraphic relations suggest that the Crystal Pass is absent and the Dawn Limestone rests unconformably on the Guilmette Formation, or that Crystal Pass equivalent rocks are present and are represented by a different facies. We tentatively assign these Crystal Pass equivalent rocks to the Guilmette Formation (fig. 2). Poole and Sandberg (1991) interpreted a near-shore, lagoonal depositional setting for the Crystal Pass Limestone in the region.

In the northwest Spring Mountains (WPT), rocks of the Lower Mississippian Narrow Canyon Formation are below the Monte Cristo Group and above the Guilmette Formation, and the Crystal Pass Limestone is presumably absent. Rocks of the Narrow Canyon Formation are part of the Mississippian central facies belt of Stevens and others (1996), a more western facies compared with the southeastern belt represented by the Monte Cristo Group in most of the quadrangle. The type area of the Narrow Canyon Formation and associated Mississippian central facies rocks is in the Spotted Range area, northwest of the Las Vegas quadrangle (Stevens and others,

1996). Although the Narrow Canyon Formation is middle Kinderhookian in age, it typically contains reworked early to middle Famennian conodonts (Harris and others, in press).

Rocks of the Lower and Upper Mississippian Monte Cristo Group are widely distributed throughout the quadrangle. These rocks are composed mostly of gray limestone deposited on an extensive carbonate platform across southern Nevada. Stevens and others (1996) defined these rocks to represent the southeastern facies belt of the southern Nevada and east-central California region. The rocks are typically bioclastic wackestone, packstone, and grainstone with abundant crinoids, solitary and colonial corals, and brachiopods. Most beds of the Monte Cristo Group are massive, but parts of the Anchor Limestone form slopes. The Monte Cristo Group maintains a thickness of about 300 m across the quadrangle, although the maximum thickness of the group is 480 m in the southern Las Vegas Range (Lundstrom and others, 1998).

The Dawn Limestone is the lowermost formation of the Monte Cristo Group. It is a dark-gray, abundantly fossiliferous limestone that overlies and markedly contrasts with the light-gray micrite of the Crystal Pass Limestone. Conodonts indicate the age of the Dawn is Kinderhookian and early Osagean and suggest that the contact with the underlying Crystal Pass Limestone ranges from conformable to disconformable (Harris and others, in press). The Anchor Limestone of early to late Osagean age (Harris and others, in press; Stevens and others, 1996) overlies the Dawn Limestone. Unlike other formations in the group, the Anchor contains 10–40 percent chert beds, some of which are spiculitic. The formation most likely represents slope deposition off a carbonate shelf margin during continued deepening (Poole and Sandberg, 1991). The succeeding Bullion Limestone is mostly massive encrinitic grainstone indicating progradation of a carbonate shelf westward and above the slope deposits of the Anchor. Conodonts indicate that much of the Anchor and Bullion Limestones in the Las Vegas Range (VT) formed during the middle Osagean, but near Indian Springs (WPT), just north of the Las Vegas quadrangle, conodonts indicate boundaries of Bullion Limestone are slightly younger, from late Osagean to early Meramecian (Stevens and others, 1996). The base of the Bullion in the WPT is also late Osagean (fig. 2). The Yellowpine Limestone, the uppermost formation of the Monte Cristo Group, yields late Osagean and Meramecian conodonts (Harris and others, in press). The Yellowpine contains conspicuously large solitary rugose corals in the Las Vegas Range (VT) and eastern Spring Mountains (KT).

The Mississippian Battleship Wash Formation locally overlies the Yellowpine Limestone of the Monte Cristo Group in the southern Las Vegas Range. Webster (1969) noted a thin Battleship Wash Formation in the Lee Canyon area of the Spring Mountains. The Battleship Wash Formation is late Meramecian to early Chesterian based on conodonts collected in the southern Las Vegas Range (Harris and others, in press). The unit is a shallow marine deposit that commonly has spirifer brachiopods and corals.

The Mississippian Indian Springs Formation of Webster and Lane (1967) was combined with the Bird Spring Formation and Callville Limestone, and it represents the basal part of these map units in the Las Vegas quadrangle. The Indian Springs Formation is Chesterian in age (Webster, 1969), and it is a good marker unit containing the biostratigraphically diagnostic *Rhipodomella nevadensis* and *Stigmaria*, as well as distinctive red and black shale beds. The formation represents deposition in a near-shore, marginal marine environment. Throughout the quadrangle, the Indian Springs Formation disconformably overlies the Monte Cristo Group or the Battleship Wash Formation and conformaby underlies the Bird Spring Formation.

Pennsylvanian and Lower Permian rocks in the quadrangle include part of the Bird Spring Formation and equivalent Callville Limestone, succeeding Lower Permian redbeds, and the Lower Permian Kaibab and Toroweap Formations. The Bird Spring Formation is widely exposed in the Las Vegas quadrangle and reaches a maximum thickness of 2,500 m in the Las Vegas Range (Lundstrom and others, 1998). The Bird Spring Formation includes carbonate shelf, slope, and basinal deposits.

We subdivided the Bird Spring Formation into two separate map units in the Las Vegas Range (VT) and in the northern Spring Mountains (LMCT and DCT). This subdivision allowed us to map important structures previously overlooked in the Las Vegas quadrangle. One of these structures is the Valley thrust fault and associated folds in the Las Vegas Range (Lundstrom and others, 1998). The lower unit of the Bird Spring Formation includes carbonate-shelf rocks of mostly Pennsylvanian age, and the upper member includes Lower Permian, uppermost Wolfcampian to Leonardian beds. The lower unit of the Bird Spring Formation is about 1,000 m thick in the quadrangle and consists of silty limestone with subordinate dolostone, siltstone, shale, and chert; these rocks are abundantly fossiliferous.

The upper unit of the Bird Spring Formation marks a significant change in depositional regime from a shelf depositional setting that typifies most of the Bird Spring in the region to a deeper water basinal and carbonate slope depositional regime. These rocks were deposited in the Bird Spring basin (Bissell, 1974), the dominant basin in the southern Nevada region that resulted from a major late Wolfcampian marine transgression and associated tectonism (Page, 1993, 1998). During this time, equivalent basin and slope facies rocks accumulated in the Dry Mountain Trough, a similar sedimentary basin in east-central Nevada (Gallegos and others, 1991; Stevens, 1991). During Leonardian time, carbonate-shelf deposits gradually prograded over the upper Wolfcampian slope to basin deposits of the Bird Spring Formation, reflecting a regional marine regression and filling of the Bird Spring basin.

Deeper water carbonate slope and basinal facies characteristic of the upper member of the Bird Spring Formation are absent in the eastern Spring Mountains (KT) and northern Bird Spring Range (BST) and are cratonic platform and inner shelf

deposits resembling the Callville Limestone in Frenchman Mountain area. For example, evaporites are in the upper part of the Bird Spring Formation 4 km south of Blue Diamond (BST) near the southern edge of quadrangle.

Strata overlying the Bird Spring Formation and Callville Limestone include the Lower Permian redbeds that reflect a significant shift from dominantly marine to mostly continental deposition. The Lower Permian redbeds are the basal sediments of the upper Paleozoic and Mesozoic continental platform assemblage (Dickinson, 1992) in the southern Nevada region. These rocks are mostly cross-bedded sandstone and subordinate siltstone, shale, conglomerate, and gypsum. Carr and others (2000) and McDonnell-Canan and others (2000) used the name Hermit Formation for these rocks in the Blue Diamond area from correlation with the Hermit Shale in the Grand Canyon region. At Frenchman Mountain, Longwell and others (1965) suggested correlation of the Lower Permian redbeds with parts of the Queantoweap Sandstone, the Supai Group, and the Coconino Sandstone—units defined in the Colorado Plateau region. The absence of fossils in these rocks, however, prevents a more definitive correlation.

The Permian Kaibab and Toroweap Formations were mapped here as one unit and are similar to units exposed in the Colorado Plateau region. Formation members defined in the plateau region were recognized in the Blue Diamond area by Carr (1992), Carr and others, (2000), Carr and McDonnell-Canan (1992), and McDonnell-Canan and others (2000). The Kaibab and Toroweap Formations form two prominent limestone ridges above the less resistant Lower Permian redbeds, and they contain abundant chert layers as well as subordinate dolostone, siltstone, shale, and gypsum. Rocks of the Kaibab and Toroweap Formations formed in shallow marine shelf to marginal marine environments. The Kaibab Formation contains nearly pure bedded gypsum in the upper Harrisburg Member, and it is locally mined and processed at the James Hardee Mine, at the top of Blue Diamond Hill in the southern part of the quadrangle. Exposures of the Kaibab and Toroweap Formations in the KT represent their westernmost limit.

Mesozoic Rocks

Mesozoic rocks in the quadrangle include the Triassic Moenkopi and Chinle Formations and the overlying Jurassic Aztec Sandstone. Most of these rocks are terrigenous clastic beds with subordinate marine rocks. Mesozoic formations in the quadrangle have been correlated with those in the Colorado Plateau region, but some members are absent because of stratigraphic thinning along major unconformities (Marzolf, 1990). The Moenkopi Formation and Aztec Sandstone have been correlated with rocks in the Jurassic arc terrane in the Mojave Desert of southeastern California (Marzolf, 1990).

The Timpoweap Conglomerate Member at the base of the Moenkopi Formation unconformably overlies rocks of the underlying Kaibab and Toroweap Formations, but in parts of the eastern Spring Mountains, it directly overlies the Lower

Permian redbeds. Rocks in the lower part of the Moenkopi Formation above the Timpoweap Conglomerate Member are mostly terrigenous sandstone, siltstone, shale, and gypsum, with minor marine carbonate rocks. These rocks are overlain by the Virgin Limestone Member, a shallow marine deposit containing pelecypods, brachiopods, and pelmatozoans. Conodonts from the Virgin Limestone Member in the Blue Diamond area are Early Triassic in age (Harris and others, in press). Marine limestone in the Virgin Limestone Member represents the final stages of marine incursion across southern Nevada (Longwell and others, 1965). The Virgin Limestone Member of the Moenkopi Formation is overlain by an upper member composed of reddish-brown claystone, siltstone, sandstone, and subordinate limestone and gypsum. Hewett (1931) reported volcanic clasts and ashy beds in the member south of the quadrangle, just west of Goodsprings, Nevada.

The Upper Triassic Chinle Formation rests unconformably on the Moenkopi Formation in the quadrangle. The Chinle consists of terrigenous sandstone, siltstone, shale, and conglomerate; no marine beds have been recognized in the formation in the southern Nevada area (Longwell and others, 1965). The erosional base of the formation is defined by the Shinarump Member, which contains distinctive clasts of reddish- to yellowish-brown chert. Volcanic clasts have been reported in the Shinarump Member in the Spring Mountains, and volcanic-derived bentonitic clays are in other parts of the unit in the region (Longwell and others, 1965). Above the Shinarump Member, the Petrified Forest Member of the Chinle Formation is recognized in the southern Nevada region; these rocks are unconformably overlain by strata equivalent with parts of the Moenave and Kayenta Formations (Wilson and Stewart, 1967; Marzolf, 1990). In this report, the Moenave and Kayenta Formation equivalent rocks are included in the upper part of the Chinle Formation.

The Jurassic Aztec Sandstone conformably overlies the Chinle Formation in the quadrangle. The Aztec is equivalent to the Navajo Sandstone in southwest Utah. The formation consists predominantly of cross-bedded sandstone; quartz grains are fine to medium grained. The Aztec Sandstone is magnificently exposed in the Wilson Cliffs in the eastern Spring Mountains, where it is truncated by the Keystone thrust fault.

The conglomerate of Brownstone Basin (Axen, 1985) may represent the only Cretaceous unit in the quadrangle; the absence of datable material within the unit has prevented determination of its age. The conglomerate of Brownstone Basin is too thin to show on map A, but its distribution is shown by Axen (1985) in the La Madre Mountain area. The conglomerate of Brownstone Basin unconformably overlies the Jurassic Aztec Sandstone locally in the La Madre Mountain area north of the La Madre fault, and in the Wilson Cliffs area south of the La Madre fault; like the Aztec Sandstone, the unit is truncated by both the Red Spring and Keystone thrusts (Davis, 1973). Axen (1985) interpreted the conglomerate of Brownstone Basin as a synorogenic deposit that formed in frontal parts of the Red Spring thrust fault. Burchfiel and Davis (1988) also reported small outcrops of the unit below

the Keystone thrust in the Wilson Cliffs area. The unit is from 2 to 5 m thick locally and consists of mixed Paleozoic and Mesozoic clasts in a matrix of red sandstone derived mostly from the Aztec Sandstone. Axen (1985) correlated the conglomerate of Brownstone Basin with the Lavinia Wash sequence, a similar synorogenic deposit exposed south of the quadrangle in the Goodsprings area (Carr, 1980; Fleck and Carr, 1990). The conglomerate of Brownstone Basin may also correlate with part of the Baseline Sandstone in the Muddy Mountains, a synorogenic deposit in the lower plate of the Muddy Mountain thrust, an equivalent to the Keystone thrust.

Tertiary Rocks

Tertiary rocks in the quadrangle include the Horse Spring Formation (Longwell and others, 1965; Bohannon, 1984), volcanic rocks in the McCullough Mountains, and the Muddy Creek Formation. The Horse Spring Formation is Miocene to Oligocene in age, volcanic rocks in the quadrangle are middle Miocene, and the Muddy Creek Formation is Pliocene to late Miocene. Tertiary rocks are significant because they provide relative age constraints for Tertiary extensional structures in the quadrangle.

Rocks equivalent to the Horse Spring Formation are exposed in several shallow basins in the southern Las Vegas Range. These rocks are deformed and are considered "syntectonic" in terms of regional Tertiary extension. Correlation of these basin-fill deposits in the quadrangle with the Horse Spring Formation is controversial. Some workers (Guth and others, 1988) argue that these rocks were deposited into uniquely separate basins from those of the Horse Spring Formation in the Lake Mead area (Bohannon, 1984). We follow Maldonado and Schmidt (1991) and Longwell and others (1965) and use the name Horse Spring Formation for these deposits as well because of similar depositional age and the fact that these deposits collectively help to constrain the age of Cenozoic extension in the region.

The Gass Peak basin (Guth and others, 1988) lies between Fossil Ridge and Gass Peak, two east- to northeast-trending ranges oroflexurally bent during movement along the Las Vegas Valley shear zone. Guth and others (1988) informally named the deposits filling Gass Peak basin the "Gass Peak formation." Maldonado and Schmidt (1991) correlated these deposits with the Horse Spring Formation and subdivided the unit into several members. The deposits consist of alluvial gravel, megabreccia, lacustrine sediments, and some volcanic rocks. K-Ar ages from interbedded volcanic rocks of the Horse Spring Formation in the Gass Peak basin range from 12 to 16 Ma (Guth and others, 1988; Maldonado and Schmidt, 1991). The deposits are tilted and faulted against Proterozoic to lower Paleozoic rocks on the flanks of Fossil Ridge and Gass Peak. The Horse Spring Formation is also exposed in a small basin in the northeastern corner of the quadrangle. Here,

conglomerates and some lacustrine beds and several beds of alkali basalt are tilted about 15–20° to the east along the Hidden Valley fault.

Andesitic lava flows were deposited in the southeastern part of the map and were correlated with the Mount Davis and the Patsy Mine Volcanics (Bingler, 1977; Anderson, 1977). These rocks have their sources in the McCullough Range (Smith and others, 1988) and Eldorado Mountains (Anderson, 1971) which are in the northern part of the Lower Colorado River extensional corridor (Faulds and others, 1994). The rocks were erupted during periods of maximum extension in the region, and at some locations in the Eldorado Mountains, they are tilted as much as 90° (Feurbach and others, 1993).

Deposits of the Muddy Creek Formation are preserved at several localities on the southwest flank of Frenchman Mountain. These sediments are part of more extensive exposures of the formation near Frenchman Mountain in the adjacent Lake Mead 1:100,000-scale quadrangle. These rocks are partly gypsiferous, red and gray sandstone and mudstone. Most of the sediments of the Muddy Creek in the region are lacustrine and are interbedded with alluvial gravels deposited mostly along Cenozoic basin margins. The Muddy Creek Formation is generally undeformed compared with underlying sediments of the Horse Spring Formation, and Longwell (1974) concluded that movement on the Las Vegas Valley shear zone must have ended by the onset of Muddy Creek deposition because the shear zone apparently does not cut these rocks.

Quaternary

To evaluate the antiquity of early man in the area, Haynes (1967) described and established the Quaternary stratigraphy (units A through G) in the Tule Springs area of the northern Las Vegas Valley. This work has remained a sound basis for subsequent Quaternary studies and mapping in the area, including this map. Though Haynes (1967) recognized paleospring point discharge associated with stratigraphic units, he interpreted some of the deposits as lacustrine, as Longwell and others (1965) had done earlier. In a Statewide evaluation of pluvial lakes and climate, Mifflin and Wheat (1979) showed that characteristics of valleys with pluvial shorelines contrasted with those of valleys with evidence of past groundwater discharge without pluvial lakes, and inferred that most Southern Nevada valleys were in the latter category. Through studies of facies relationships and various paleoenvironmental data, Quade (1986) established the association of the fine-grained units with expanded regional ground-water discharge in the Las Vegas Valley during late Quaternary climates that were significantly cooler and moister than present. Later studies (Quade and Pratt, 1989; Quade and others, 1995, 1998) have expanded the paleoenvironmental, geochemical, and geochronologic databases to show that late Pleistocene ground-water discharge recurred episodically in many valleys of the region in response to climate change and its effects on ground-water systems.

This 1:100,000-scale geologic map, which synthesizes recent geologic maps of 7.5' quadrangles (Bell and others, 1998, 1999; dePolo and others, 1999; Lundstrom and others, 1998, 2003), shows the extent of fine-grained deposits and the relationships of these deposits to faults, alluvial fan gravel, bedrock geology, and physiography. There is a common spatial association of the fine-grained deposits with Quaternary faults that offset these deposits. Faults that juxtapose strata of differing transmissivities can form hydrologic barriers that increase surficial discharge of shallow ground water.

Extensive areas of fine-grained deposits dominate the surface of lower parts of the Las Vegas and Pahrump Valleys. Most of these deposits were formed in various types of wetlands during episodes of past ground-water discharge that were much larger than historically observed. The most areally extensive period of discharge during the past 50,000 years occurred between about 40 ka and 25 ka when map unit **Qscd** was deposited, preceding and overlapping the last global glacial maximum. Comparison to proxy climate records (for example, Spaulding, 1990) indicates that this was a period of relatively high effective moisture and recharge. A less extensive period of past discharge occurred during 13–8 ka, when unit **Qse** was deposited.

Extensive coalescing alluvial fans exist between the fine-grained areas of past discharge and the upland sources of the fans. Like past discharge, fan sedimentation was also episodic and hydroclimatically driven. Extensive Holocene alluvial-fan sedimentation (**Qay**, especially **Qayo**) overlapped the younger period of prehistoric ground-water discharge at about 13–8 ka (Liu and others, 2000). Somewhat wetter than historic climates with surface flood events of larger than historic magnitude produced these fans. In contrast, the most extensive period of past discharge (40–25 ka) did not coincide with fan-building, but overlapped and preceded major fanhead incision (**Qaiy**) in watersheds that included the highest parts of the Spring Mountains. The fanhead incision required concentrated runoff and low sediment yield, and was probably dominated by snowmelt. The latter part of the last major episode of carbonate cementation in soils in the area occurred during this period of enhanced late Pleistocene effective moisture. The carbonate cementation from this period helps to distinguish well-cemented Pleistocene alluvial soils from non-cemented Holocene soils. Prior to fanhead incision, the penultimate period of extensive fan deposition (**Qai**) and aggradation similar to that in the Holocene occurred during 120–50 ka, probably in response to high-intensity rainfall, runoff, and erosion of uplands. Similar cycles of fan building and ground-water discharge probably occurred during the early and middle Pleistocene and perhaps earlier, but deposits including **Qao**, **QTa**, and **QTs** are not dated nor exposed well enough to test such speculation.

Structure of the Las Vegas 30′ × 60′ Quadrangle

Introduction

The first major structural episode that affected the rocks in the quadrangle was the development of thrust faults and folds related to the Sevier orogeny in Cretaceous time. Major, southeast-directed, Mesozoic thrust faults are exposed in the Spring Mountains and Sheep and Las Vegas Ranges; they are, from west to east, the Wheeler Pass and equivalent Gass Peak thrusts, the Lee Canyon and Macks Canyon thrusts and equivalent Valley thrust, the Deer Creek thrust, the Keystone thrust, and the Bird Spring thrust (fig. 3*A*, map A). Burchfiel and others (1974) reported 36–75 km of shortening across the Spring Mountains as a result of Mesozoic thrusting.

Late Tertiary extension in the Las Vegas quadrangle resulted in regional strike-slip and normal faulting, landsliding, and volcanism. The major extensional structures in the quadrangle include the dextral Las Vegas Valley shear zone and the State Line fault zone (fig. 3*A*). Other structures that formed during extension include high-angle normal and strike-slip faults and landslide deposits shed off the Spring Mountains.

Quaternary faults are exposed in both Pahrump and Las Vegas Valleys, where they are generally associated with paleo-spring discharge deposits of Pleistocene age. These faults are discontinuous along strike, and have low slip rates. The origin of these faults is controversial, and recent studies suggest that many of these faults are indeed tectonic features.

Mesozoic Thrust Faults

Spring Mountains

Wheeler Pass Thrust

The Wheeler Pass thrust (fig. 3*A*) juxtaposes Late Proterozoic and Lower Cambrian clastic and carbonate units in the upper plate against mostly overturned Permian beds of the Bird Spring Formation in the lower plate. Stratigraphic separation on the thrust is approximately 4.8 km (Vincelette, 1964), and Fleck (1970) estimated about 7 km of horizontal shortening. This southeast-vergent thrust strikes northeast like other major thrusts in the quadrangle but dips more steeply (as much as 50°) (Nolan, 1929; Vincelette, 1964; Burchfiel and others, 1974). The Wheeler Pass thrust and parts of the equivalent Gass Peak thrust, as well as the Valley thrust, are technically reverse faults, but we refer to these as "thrusts" to maintain consistency with past publications that discuss these

structures. Along the southern segment of the Wheeler Pass thrust (southwest of the Trough Spring fault), upper-plate rocks of the Wood Canyon, Carrara, and Bonanza King Formations are folded into a broad north-northeast-trending syncline (Wheeler syncline of Nolan, 1924). The thrust here is cut by several northwest-striking high-angle faults, the most prominent being the Trough Spring fault (map A). The central fault segment (northeast of the Trough Spring fault) has upper-plate rocks composed of Stirling Quartzite and Wood Canyon Formation (cross section *C–C′*, map sheet 1) deformed by folds and associated small-offset imbricate thrusts. The northern fault segment is concealed by Tertiary and Quaternary basin-fill deposits south of Indian Ridge (map A).

About 15 km northwest of the Wheeler Pass thrust in the northwest part of the quadrangle, the Stirling Mine fault (map A) strikes northeast and, although its trace is concealed by Quaternary alluvium, the fault separates Lower Cambrian and Late Proterozoic rocks on the northwest side from Cambrian and Ordovician rocks on the southeast side. Stratigraphic throw on the fault was estimated at about 1.5 km (Burchfiel and others, 1974). Snow (1992) proposed that the Stirling Mine fault is a segment of a hypothetical thrust he called the Kwichup thrust, which is at a higher structural level than the Wheeler Pass thrust. Detailed mapping of contractile structures in the block northwest of the fault by Abolins (1999) supports the existence of the Kwichup thrust. Snow (1992) interpreted that other parts of the Kwichup thrust plate were excised by normal faults and concealed in the subsurface.

Lee Canyon and Macks Canyon Thrusts

The Macks Canyon and Lee Canyon thrusts (fig. 3*A*) are discussed together because the Macks Canyon thrust is an upper plate splay of the Lee Canyon thrust (Burchfiel and others, 1974). Both faults strike northeast and lose displacement in that direction. Southwest of the La Madre fault (fig. 3*A*), the Macks Canyon thrust has a small amount of displacement compared to the Lee Canyon thrust, offsetting chiefly Cambrian and Ordovician rocks, and it terminates to the south against the Trough Spring fault where it most likely is concealed in the subsurface. Northeast of the La Madre fault, upper-plate Mississippian rocks are in fault contact with mostly Pennsylvanian rocks in the lower plate. Here, the thrust has utilized the shaly Indian Springs Formation (Upper Mississippian) as a glide plane.

Fleck (1970) estimated maximum stratigraphic separation on the Lee Canyon thrust at 4.8 km and the amount of horizontal shortening at about 7 km. Along the southern segment of the Lee Canyon thrust, carbonates of the Bonanza King Formation in the upper plate are juxtaposed against mostly west-dipping rocks of the Bird Spring Formation in the lower plate. The central fault segment in upper Lee Canyon has upper plate rocks consisting mostly of Upper Cambrian Nopah Formation above lower plate rocks ranging from Ordovician to Pennsylvanian. These rocks are highly faulted by local imbricate thrusts that are in turn cut by northwest-striking

high-angle faults. Northeast of the La Madre fault, Ordovician rocks in the upper plate are juxtaposed against overturned beds of the Bird Spring Formation in the lower plate. At the northernmost exposures of the thrust in lower Lee Canyon, like the Macks Canyon thrust, Mississippian rocks in the upper plate are juxtaposed against the Bird Spring Formation in the lower plate. We interpret the two thrusts to merge in this area (fig. 3*A*).

Deer Creek Thrust

Fleck (1970) estimated a maximum stratigraphic separation on the Deer Creek thrust of about 3 km and lateral shortening of about 5.3 km. The southern segment of the Deer Creek thrust is bounded on the north by the La Madre fault, and on the south by the Griffith fault (fig. 3*A*, map A). Along this segment, the thrust juxtaposes Nopah Formation in the upper plate against thrust-imbricated Mississippian and Pennsylvanian rocks in the lower plate. The central segment of the fault is exposed in upper Lucky Strike Canyon, where rocks of the Nopah Formation in the upper plate are juxtaposed against rocks of the Bird Spring Formation in the lower plate (map A). The northern segment of the fault is exposed in lower Lucky Strike Canyon, where the fault juxtaposes Nopah Formation in the upper plate against Mississippian rocks in the lower plate.

The Kyle Canyon thrust is exposed about 3.5 km east of the southern Deer Creek thrust segment where it offsets only rocks of the Bird Spring Formation (cross section *C–C'*, map sheet 1). This thrust is probably a frontal splay of the Deer Creek thrust because both faults are in the same structural block and terminate at the Griffith fault. Fleck (1970) estimated maximum stratigraphic separation on the thrust to be about 0.9 km, and lateral shortening about 0.8 km.

Keystone and Red Spring Thrusts

The Keystone thrust is exposed for nearly 70 km along the length of the eastern Spring Mountains. It is one of the best exposed thrust faults in the entire Cordillera, and forms the eastern limit of the Cordilleran fold and thrust belt in the region. We follow Longwell and others (1965) and use the name Keystone thrust for the spectacularly exposed thrust above the Wilson Cliffs, between the La Madre and Cottonwood faults (fig. 3*A*; map A), although Burchfiel and others (1997) alternatively refer to this thrust as the Wilson Cliffs thrust. The Keystone thrust juxtaposes gray carbonate rocks of the Cambrian Bonanza King Formation in the upper plate against red- and tan-weathering beds of the Jurassic Aztec Sandstone in the lower plate. The thrust is well exposed in Red Rock Canyon, just south of the La Madre fault. Here, the thrust plane is in a ramping configuration (dipping 30° to the northwest) above an overturned footwall syncline in Triassic and Jurassic rocks (cross section *C–C′*, map sheet 1). Burchfiel and others (1974) estimated values ranging from 11.6 to 23 km horizontal displacement along the Keystone thrust based on several respective models they considered for the thrust.

Near the southern boundary of the quadrangle, the Keystone thrust is offset by several high-angle faults. The most prominent is the Cottonwood fault, a northwest-striking fault that forms the southern boundary of the Wilson Cliffs (fig. 3A). The Cottonwood fault accommodated down-to-the-southwest displacement, and juxtaposed Paleozoic rocks in the downthrown block against Aztec Sandstone in the upthrown block. In contrast to this large stratigraphic throw, the fault only offsets the Keystone thrust about 60 m (Davis, 1973).

The La Madre fault cuts the Keystone thrust at the north end of Wilson Cliffs. Northeast of the La Madre fault in the La Madre Mountains area, the Keystone thrust is in fault contact with the Red Spring blocks (fig. 3A). The Red Spring blocks are three northeast tilted structural blocks bound by the Keystone thrust on the northwest, and by the La Madre, Turtlehead Mountain, Brownstone Basin, and Box Canyon faults, from southeast to northeast. Between the La Madre and Brownstone Basin faults, the Keystone thrust is at the boundary between the northeast-striking beds of La Madre Mountain (upper plate) and the northwest-striking beds of the southern two Red Spring blocks (lower plate) (Axen, 1985). Axen (1985) interpreted the Keystone thrust northeast of the Brownstone Basin fault to cut up-section and juxtapose rocks as young as the Upper Cambrian Nopah Formation (upper plate) against a discordant Cambrian through Mississippian sequence (lower plate) in the northernmost Red Spring block (map A). The upper plate rocks are highly deformed by small imbricate thrusts, tear faults, and folds. Several smaller high-angle faults offset the Keystone thrust above the northern Red Spring blocks.

The northwest-striking Red Spring thrust is exposed repeatedly in all three Red Spring blocks (fig. 3A). The thrust dips from 35° to 40° to the northeast, and like the Keystone thrust, juxtaposes upper-plate rocks of Cambrian Bonanza King Formation against Jurassic Aztec Sandstone in the lower plate (map A). The structural relationship between the Keystone and Red Spring thrusts is controversial because of complex structural relations. The controversy is long-lived, dating to the 1920s when Longwell (1926) first interpreted the Red Spring thrust as a separate thrust that pre-dated the Keystone thrust and was subsequently overridden by the younger Keystone thrust. Longwell (1960) later reinterpreted the Red Spring thrust to be equivalent with the Keystone thrust, and that the Red Spring blocks were fragments of the Keystone plate broken off the frontal edge and overridden by the advancing Keystone thrust. Several structural models for these thrusts evolved from Longwell's interpretations.

Model 1, shown in figure 3B, is similar to Longwell's 1926 interpretation and assumes that the Red Spring thrust pre-dates and was overridden by the younger Keystone thrust (Axen, 1985; Davis, 1973; Burchfiel and Davis, 1988; Burchfiel and Royden, 1984; Burchfiel and others, 1997). Proponents of the model further assume that the Wilson Cliffs thrust and the Cottonwood thrust in the Goodsprings area correlate with the Red Spring thrust, and were all overridden by the younger Keystone thrust. The fault traditionally shown on other published maps as the Keystone thrust (Longwell, 1926; Secor, 1963; Burchfiel and others, 1974) above the Wilson Cliffs (fig. 3A) was named the Wilson Cliffs thrust by Burchfiel and Royden (1984) and Burchfiel and others (1997) (fig. 3B). They mapped the Keystone thrust at yet a higher structural level to the west, where in the Mountain Springs area they demonstrated that the fault was continuous and similar in style to the type Keystone thrust of the Goodsprings area. In the Mountain Springs area, Burchfiel and others (1997) distinguished the Keystone and Wilson Cliffs thrusts based on internal structure and detailed mapping of the Bonanza King Formation. They determined that the rocks in the Keystone plate are less deformed than in the Wilson Cliffs plate. The Keystone thrust juxtaposes relatively undeformed older units against younger units of the Bonanza King Formation that are highly deformed by imbricate thrusts and isoclinal folds.

Model 1 is based mostly on (1) the presence of high-angle faults (La Madre and Cottonwood faults) that cut and accommodate significant offset of the Red Spring, Wilson Cliffs, and Contact thrust plates but fail to appreciably offset the Keystone thrust and (2) the fact that the Red Spring and Contact thrusts lie topographically below the Keystone plate. Recognition of pre-Keystone high-angle faults led Davis (1973) and Burchfiel and Davis (1988) to propose that the Red Spring, Wilson Cliffs, and Contact thrusts were emplaced, faulted by northwest-striking high-angle faults, such as the Cottonwood fault and faults that bound the three Red Spring blocks, and then the part of the thrust plate between the La Madre and Cottonwood faults was uplifted and eroded away. These events were followed by emplacement of the younger Keystone thrust.

Model 2 (fig. 3C) is adapted from Longwell's 1960 interpretation of the Keystone and Red Spring thrusts, and requires only one episode of thrusting. Longwell (1960) proposed that the Red Spring and Keystone thrusts are equivalent and that the Red Spring thrust plates were parts of the Keystone plate that were broken and downdropped from frontal parts of the thrust plate, then overridden by the advancing Keystone thrust. Matthews (1988, 1989) also believed that the Keystone and Red Spring thrusts are equivalent, but he disagreed with Longwell's "overriding" concept and proposed an alternative model for the thrust faults in the Red Spring blocks (fig. 3C). Although the Red Spring blocks appear to lie structurally below the Keystone thrust, Matthews (1988) argued that the blocks stand topographically above the Keystone thrust and were displaced from the Keystone plate through a combination of vertical and horizontal axis rotation. He believed that this deformation occurred during oroclinal bending caused by Miocene movement on the Las Vegas Valley shear zone. In the southern two Red Spring blocks, Matthews (1989) interpreted the Keystone thrust segments (as shown on map A) as south-dipping normal faults that downdropped and rotated the blocks from La Madre Mountain to conceal the trace of the true Keystone thrust (fig. 3C). He interpreted the mapped

Keystone thrust in the northernmost Red Spring block (map A) as a Cenozoic thrust that formed as a result of extreme shortening accommodating oroclinal bending of the block by the Las Vegas Valley shear zone.

Matthews (1989) considered the Wilson Cliffs thrust of Burchfiel and Royden (1984) and Burchfiel and others (1997) to be the traditionally mapped Keystone thrust, and the thrust Burchfiel and Royden (1984) and Burchfiel and others (1997) identified as the Keystone thrust, as part of a deformed zone in the upper plate. Although Davis (1973), Axen (1985), and Burchfiel and Davis (1988) correlated the Contact thrust in the Goodsprings area with the proposed pre-Keystone Red Spring and Wilson Cliffs thrusts, Fleck and Carr (1990) and Matthews (1988) interpreted the Contact and Keystone thrusts as contemporaneous structures. The Cottonwood fault was interpreted by Matthews (1988) as connecting with the Contact thrust to form a lateral ramp of a lower plate duplex zone of the Keystone thrust, rather than a pre-Keystone age high-angle fault.

Longwell's (1960) interpretation and model 2 are appealing because of their simple view of the thrusts along the eastern Spring Mountains as one integrated system of the same age. Even Longwell (1960) commented that identical rocks in the Red Spring and Keystone thrust plates implied their equivalence, and Fleck and Carr (1990) concluded coeval emplacement of the Contact and Keystone thrusts in the Goodsprings area. Nevertheless, thrust models in the La Madre Mountain area and Red Spring blocks remain equivocal and need further evaluation.

Bird Spring Thrust

The Bird Spring thrust is the structurally lowest thrust in the quadrangle (fig. 3A, cross section C–C', map sheet 1). Its trace is subparallel to the Keystone thrust, and it extends northward more than 30 km from the southern Bird Spring Range to the Blue Diamond area in the southern part of the quadrangle; its throw decreases northward along strike. At its southernmost exposures, south of the quadrangle in the Goodsprings area, rocks of the Bonanza King Formation in the upper plate are juxtaposed against lower-plate rocks of the Aztec Sandstone (Burchfiel and Davis, 1988). In the quadrangle, along the east side of Blue Diamond Hill, upper-plate Permian redbeds are juxtaposed against lower-plate rocks of the Virgin Limestone Member and upper red member of the Triassic Moenkopi Formation (cross section C–C', map sheet 1). Beds of the Virgin Limestone Member are locally highly folded at the base of the thrust, especially just north of State Highway 160. The thrust appears to terminate at the La Madre fault. However, new mapping for this report indicates that it continues northward across the fault and is concealed by a shallow landslide block composed of the Kaibab and Toroweap Formations (see landslide section below).

Sheep Range and Las Vegas Range

Gass Peak and Valley Thrusts

Major Mesozoic structures in the Sheep and Las Vegas Ranges are the Gass Peak thrust and the Valley thrust and structurally related overturned limbs of an anticline-syncline pair (fig. 3A, map A, cross sections A–A' and B–B', map sheet 1). Strike of these structures changes from east-west near the margin of Las Vegas Valley to northeast in the northern part of the quadrangle. This pattern resulted from oroclinal bending along the Miocene right-lateral Las Vegas Valley shear zone. Nelson and Jones (1987) documented more than 50° of clockwise, vertical-axis rotation of rocks (Cambrian Bonanza King Formation) in the Las Vegas Range, near the shear zone. They concluded that surface deformation resulting from oroclinal bending is characterized by small fault-bounded blocks (less than 5 km in length) that rotated independently due to ductile deformation related to the shear zone at depth. Lundstrom and others (1998) detailed this style of deformation in their mapping of Paleozoic rocks adjacent to the shear zone.

The Gass Peak thrust has Proterozoic and Cambrian rocks in the upper plate and Lower Permian (Leonardian) rocks in the lower plate (Guth, 1990; Maldonado and Schmidt, 1991; Ebanks, 1965). Guth (1981) reported the thrust to have as much as 5.9 km of stratigraphic separation and greater than 30 km of horizontal displacement. Dips of the thrust range from 40° to 47° to the northwest, and Guth (1990) interpreted a ramp-flat-ramp geometry with a decollement in the Proterozoic Johnnie Formation. Ebanks (1965) described a complex imbrication zone along the thrust near the northern edge of the quadrangle, involving rocks of the Stirling Quartzite and Carrara Formation (map A).

Structure in the southernmost Las Vegas Range is dominated by the Valley thrust and overturned limbs of an anticline-syncline pair (fig. 3A, map A, cross sections A–A' and B–B', map sheet 1). These structures are subparallel to and in the lower plate of the Gass Peak thrust. The Valley thrust (Lundstrom and others, 1998) strikes northeast and is traced 8 km north from the margin of Las Vegas Valley, where it is at the base of Mississippian rocks that form the core of an asymmetric, southeast-verging anticline (Valley anticline) (fig. 3A, map A). The thrust placed Mississippian rocks in the upper plate above an overturned syncline (Valley syncline) in lower plate Pennsylvanian and Permian rocks.

Two segments of the Valley thrust are separated by a broad, northwest-trending reentrant in the southernmost Las Vegas Range, referred to as East Pass (fig. 3A; map A) (Ebanks, 1965). The thrust is best exposed southwest of East Pass, where it dips from 40° to 45° to the northwest and ramps along shaly beds of the Mississippian Indian Springs Formation (map A). Northeast of the pass, Mississippian rocks die out and we interpret that the thrust terminates into the Valley anticline. In the East Pass area, the Valley thrust and folds are offset subhorizontally about 1 km by

a northeast-dipping detachment fault (map A). This fault is concealed beneath Quaternary deposits in East Pass, but exposed northeast of the pass in the Bird Spring Formation, where it dips 25° to the northeast (map A). Because the fault offsets Mesozoic structures, we conclude that it formed during oroflexural bending and structural adjustment associated with movement on the Las Vegas Valley shear zone during the Miocene. Recognition of the Valley thrust and folds has led to refinement of correlation of major Mesozoic thrust faults across the shear zone (Lundstrom and others, 1998).

Correlation of Thrusts

Mesozoic thrust faults in the Sheep and Las Vegas ranges correlate with thrusts in the Spring Mountains. The thrusts were offset right-laterally along the northwest-striking Las Vegas Valley shear zone during Miocene time and indicate 50 km of horizontal displacement (Duebendorfer and Black, 1992; Longwell, 1960; Wernicke and others, 1988). Of these thrusts, the Gass Peak thrust in the southern Sheep Range correlates with the Wheeler Pass thrust in the northern Spring Mountains (Guth, 1981, 1990; Wernicke and others, 1988; Burchfiel and Davis, 1988). The Wheeler Pass thrust equivalent, to the south and west of the quadrangle, however, is controversial. Snow (1992) correlated the Wheeler Pass thrust with the Montgomery thrust, exposed in the Montgomery Mountains west of the quadrangle. Snow (1992) proposed that these thrusts were offset right-laterally during the Miocene, and estimated 20 km of right-lateral displacement between the Spring and Montgomery Mountains.

Snow (1992) correlated the Kwichup thrust (Stirling Mine fault in northwest Spring Mountains) with the Lemoigne thrust in the Cottonwood Mountains of the Death Valley region. Abolins (1999) alternatively correlated the Kwichup thrust with the Montgomery thrust, and suggested that the Wheeler Pass thrust equivalent is south of the Spring Mountains and is concealed beneath Pahrump Valley, or it may be in the Kingston Range, 40 km south of the quadrangle. This correlation is based on detailed mapping that revealed continuity of stratigraphic units between the Spring Mountains and Montgomery Mountains and similarity of structure between the Kwichup and Montgomery faults. Paleogeographic reconstruction of major thrusts in the Death Valley and Spring Mountains area indicates 85–115 km of west-northwest translation of the Cottonwood Mountains and Panamint Range with respect to the Spring Mountains (Abolins, 1999). This large-magnitude translation was accommodated in part by movement along major Tertiary extensional structures that separate the Spring Mountains from the Death Valley extensional terrain, including the State Line fault zone.

Wernicke and others (1988) correlated the Lee Canyon thrust in the Spring Mountains with the Dry Lake thrust in the Dry Lake Range northeast of Las Vegas. Marked similarities between the newly mapped Valley thrust and the Lee Canyon

and Macks Canyon thrusts along the margins of Las Vegas Valley suggest that it is more likely that these thrusts are correlative (Lundstrom and others, 1998). For example, both thrusts have ramp-anticlines composed of Mississippian rocks in their upper plates, both were thrust above overturned Pennsylvanian and Permian rocks in their lower plates, and both thrusts ramped along shaly units of the Mississippian Indian Springs Formation. In addition, the distance from the Gass Peak thrust to the Valley thrust (8–9 km) is more compatible with the distance measured from the Wheeler Pass thrust to the Lee Canyon and Macks Canyon thrusts (9–10 km) than the distance measured from the Gass Peak thrust to the Dry Lake thrust (22 km). Therefore, Lundstrom and others (1998) correlated the Dry Lake thrust with the Deer Creek thrust, 18–19 km east of the Wheeler Pass thrust in the Spring Mountains. Both thrusts have lower Paleozoic rocks (Cambrian and Ordovician) in their upper plates and contain upper Paleozoic and Mesozoic rocks in their lower plates. In addition, similarities in the Eureka Quartzite in these thrust plates (Ross, 1964) supports this correlation (see Eureka Quartzite on p. 22).

The Keystone thrust in the eastern Spring Mountains correlates with the Muddy Mountain thrust, which is east of Las Vegas in the Muddy Mountain area (Fleck, 1970; Fleck and Carr, 1990; Wernicke and others, 1988). Both thrusts form the eastern limit of Mesozoic thrusts in this part of the Cordillera, and both thrusts have Paleozoic rocks in their upper plates and Mesozoic rocks in their lower plates.

Age of Thrusts

The age of thrust faults in the quadrangle is poorly constrained, although most workers agree that the major episode of thrusting in this part of southern Nevada occurred during the Sevier orogeny in Cretaceous time (Fleck, 1970; Guth, 1990; Longwell and others, 1965). The age of the Keystone thrust in the region is based mostly on isotopic dates of synorogenic deposits cut by correlative thrust faults south (Goodsprings area) and east (Muddy Mountains) of the quadrangle. In the Goodsprings area, the Keystone and Contact thrusts both cut deposits of the Lavinia Wash sequence (Carr, 1980). Preliminary conventional K-Ar dates for the Lavinia Wash sequence were Late Jurassic to Early Cretaceous (154±10 Ma) and thus established a maximum age of emplacement for these thrusts. New laser-fusion ^{40}Ar/^{39}Ar methods yielded dates of 99±0.04 Ma for the sequence, suggesting a further refined maximum age for the thrusts as late Early Cretaceous (Fleck and Carr, 1990).

In the Muddy Mountains, the Muddy Mountain thrust (Keystone equivalent) cuts synorogenic deposits of the Baseline Sandstone of late Albian to Cenomanian(?) age (Bohannon, 1984), thus establishing a maximum age of thrust emplacement similar to the age of the Keystone and Contact thrusts in the Goodsprings area. The Baseline Sandstone

and Lavinia Wash sequence are similar in age, depositional environment, and structural position, and thus led Fleck and Carr (1990) to correlate these units along the frontal Sevier thrust belt. They further concluded that correlation of these units supports regional correlation of the Keystone and Muddy Mountain thrusts.

High-angle Normal and Strike-Slip Faults

Spring Mountains

Trough Spring Fault

The Trough Spring fault (Vincelette, 1964) is a high-angle normal and (or) oblique-slip fault that offsets the Wheeler Pass thrust just south of Wheeler Pass, hence the most recent movement on the fault post-dated thrusting, most likely during the Cenozoic (fig. 3A, map A). Along most of its trace, the fault strikes northwest, and it accommodated down-to-the-southwest displacement where upper Paleozoic rocks (mostly Bird Spring Formation) are downdropped against lower Paleozoic rocks. Vincelette (1964) estimated 1,828 m of displacement along the northwest segment of the fault. South of Clark Canyon, the fault abruptly changes strike to the northeast. Vincelette (1964) explained this curious bend by differential movement of upthrown and downthrown blocks, whereby the downthrown block rotated and tilted northeastward during faulting, independently of the relatively undeformed northwest-dipping beds in the upthrown block.

La Madre Fault

The La Madre fault (fig. 3A, map A) strikes northwest and can be traced for more than 50 km across the Spring Mountains. The fault has been proposed to have multiple histories (Axen, 1985; Davis, 1973). As suggested by Longwell and others (1965), it may have been active as a tear fault with mostly strike-slip movement during Mesozoic thrusting, similar to the Griffith fault. In the La Madre Mountain area, Axen (1985) estimated 1,130–1,230 m of vertical separation along the fault prior to emplacement of the Keystone thrust, and only 100–200 m of post- or syn-Keystone separation. Latest movement on the fault was speculated by Longwell to be related to the Las Vegas Valley shear zone. Page (new mapping, this report) observed older basin-fill deposits (map unit **QTa**) that are tilted as much as 15° to the south along the La Madre fault in the Cold Creek area, and Axen (1985) reported that Quaternary(?) colluvial deposits may be deformed by the fault in the Angel Peak area.

Strike-slip movement on the La Madre fault is implied by apparent right-lateral offset of the Keystone, Macks Canyon, and Lee Canyon thrusts in the Spring Mountains. Because the fault offsets major thrusts in the quadrangle, the latest significant movement on the fault probably occurred during Cenozoic extension. In the La Madre Mountain area, the La Madre fault is coincident with a distinct structural bend in the Keystone thrust plate (Longwell and others, 1965). It there forms a boundary separating a more deformed zone to the northeast (where the Keystone thrust is in contact with the northeast-tilted Red Spring blocks) from a less deformed zone to the southwest, where the Keystone thrust truncates the Aztec Sandstone. This pattern suggests that the La Madre fault, together with the Turtlehead Mountain, Brownstone Basin, and Box Canyon faults, accommodated oroflexural bending of the La Madre Mountain block along the Las Vegas Valley shear zone (fig. 3A).

Griffith Fault

The Griffith fault is a northwest-striking high-angle fault exposed for nearly 22 km across the central Spring Mountains (fig. 3A, map A). Along the southeastern half of the fault, highly folded Mesozoic rocks on the downthrown side (southwest) are juxtaposed against upper Paleozoic rocks of the Bird Spring Formation on the upthrown block. The upthrown block is a horst block bounded on the southeast by the Griffith fault and on the northeast by the La Madre fault. The same Mesozoic rocks are exposed northeast of the La Madre fault and southwest of the Griffith fault. In this horst block, the Deer Creek and Kyle Canyon thrusts terminate against the Griffith fault, although the fault does not offset the Keystone or Lee Canyon thrusts. This suggests that the Griffith fault was active during thrusting as a tear fault that accommodated differential movement across the fault (Burchfiel and others, 1974; Longwell and others, 1965).

Wheeler Graben

The Wheeler graben (named here) is exposed in the upper plate of the Wheeler Pass thrust in the northwest Spring Mountains (fig. 3A, map A). It is bounded by high-angle normal faults that juxtapose Devonian and Mississippian rocks in the downthrown blocks within the graben against Cambrian and Ordovician rocks in upthrown blocks outside the graben (map A; cross section C–C′). The graben has a maximum width of 7 km and it pinches out to the south, just west of Wheeler Well. It is partly asymmetric, with several major northwest- to southwest-dipping normal faults of greatest displacements (2–3 km) bounding the east side, and subordinate northeast-dipping antithetic faults with lesser displacements on the west side (cross section C–C′, map sheet 1). The graben corresponds to a broad saddle in the gravity field, with lowest gravity values occurring just east of the eastern graben margin (map B, map sheet 2). The breadth of this gravity low and its correspondence to a similar saddle in the magnetic field suggest that the graben and these geophysical anomalies are caused by deeper sources that reflect topography on the denser and more magnetic basement.

Sheep and Las Vegas Ranges

Major high-angle faults in the Sheep and Las Vegas Ranges include from west to east the Yucca Forest, Mormon Pass, Fossil Ridge, Gass Spring, and Hidden Valley faults (fig. 3*A*). The Yucca Forest and Mormon Pass faults bound Yucca Forest, a broad alluvial basin in the southern Sheep Range (map A). The basin extends northward about 8 km into the Indian Springs quadrangle, and southward it pinches out along Fossil Ridge near the margin of Las Vegas Valley. The Yucca Forest fault is mostly concealed along the eastern flank of the southern Sheep Range, but near the northern edge of the quadrangle, it juxtaposes rocks of the Mississippian Monte Cristo Group on the upthrown side with late Miocene(?) to Pleistocene alluvium (**QTa**) on the downthrown side.

The Mormon Pass fault bounds the Fossil Ridge horst block on the southeastern side of Yucca Forest. Guth (1980) mapped the fault farther north in the Indian Springs quadrangle as the eastern breakaway of the Miocene Sheep Range detachment zone. Along Fossil Ridge, the fault places Cambrian and Ordovician rocks on the upthrown side against Tertiary and Quaternary deposits as young as latest Pleistocene on the downthrown side.

The Gass Peak basin (Guth and others, 1988; Guth, 1990) (fig. 3*A*) is bound on the northwest by the Fossil Ridge fault, and on the southeast side by the Gass Spring fault; dips on the Gass Spring fault range from 55° to 70° to the northwest. Miocene deposits of the Horse Spring Formation are faulted against Proterozoic and Paleozoic units along these faults at the basin margins and are locally deformed within the basin. Gravity lows correspond with the Yucca Forest and Gass Peak basins. However, a three-dimensional inversion of gravity data indicates that these Cenozoic basins probably are no greater than 0.5 km deep (Langenheim and others, chapter B, this report; see cross section *B–B′*, map sheet 1).

The Hidden Valley fault (named here) in the Las Vegas Range parallels and, in places, offsets the Valley thrust and the Valley anticline (map A). The fault strikes northeast and dips from 55° to 60° to the northwest, and offsets rocks of the Bird Spring Formation along most of its trace. In the northeastern corner of the quadrangle (map A), the fault bounds a small Cenozoic basin where it cuts (and accommodated tilting of) deposits of the Horse Spring Formation that contain interbedded olivine basalts (**Thb**) dated at 16.4 Ma (Feuerbach and others, 1993). At this location, the hanging wall of the fault is an east-tilted half-graben.

High-angle normal faults in the Sheep and Las Vegas Ranges, such as the Mormon Pass and Hidden Valley faults (map A), appear to be oroflexurally bent along the Las Vegas Valley shear zone, implying they pre-dated major movement on the shear zone and played a role in the early development of Cenozoic extensional basins in the region from about 20 to 15 Ma. Guth and others (1988) proposed that early movement on the Mormon Pass fault induced development of the Yucca Forest and Gass Peak basins. Most high-angle normal faulting, however, probably occurred during or shortly after major movement on the Las Vegas Valley shear zone from about 14 to 8 Ma (Duebendorfer and Black, 1992), a time of prolific extensional faulting in southern Nevada.

Las Vegas Valley Shear Zone

The Las Vegas Valley shear zone (LVVSZ) (fig. 3*A*; map A) is one of the major structures in the Las Vegas quadrangle. Because it is concealed by thick Tertiary and Quaternary basin-fill deposits, its geometry and location are based primarily on geophysical studies (Langenheim and others, 1997, 1998; chapter B of this report). The LVVSZ is a northwest-striking, right-lateral strike-slip fault zone that is traced about 150 km from Mercury, Nevada, southeast to the Lake Mead area. Fault segments of the LVVSZ that break the surface are exposed in the Corn Creek Springs area (map A). The effects of the LVVSZ include oroflexural bending and offset of major Mesozoic thrust faults and folds. Offset of Mesozoic thrust faults across Las Vegas Valley indicate 48±7 km of right-lateral separation (Wernicke and others, 1988); this estimate includes bending of the Las Vegas Range. Offset decreases at the eastern and western ends of the shear zone. For this reason, the LVVSZ and associated strike-slip faults are considered transverse structures (transverse zones) which separate regions of different rates, amounts, and types of extension and magmatism (Fleck, 1970; Guth, 1981; Wernicke and others, 1982; Duebendorfer and Black, 1992; Rowley, 1998).

Paleomagnetic data (Sonder and others, 1994; Nelson and Jones, 1987) indicated a 20-km-wide zone of clockwise rotation as great as 100° in rocks as young as 13.5 Ma adjacent to the LVVSZ. The paleomagnetic data, along with other structural data, bracket the principal period of movement along the LVVSZ between 14 and 8.5 Ma (Duebendorfer and Black, 1992; Duebendorfer and Simpson, 1994). Wernicke and others (1982, 1988) interpreted that detachment faults are the primary structures in the area and the LVVSZ and Lake Mead fault system are in effect tear faults in the hanging wall of a detachment. Similarly, Duebendorfer and Wallin (1991) and Duebendorfer and Black (1992) suggested that normal faulting and strike-slip motion on the LVVSZ were coeval and thus root into the Saddle Island detachment fault, 25 km southeast of Frenchman Mountain in the adjacent Lake Mead quadrangle. The Saddle Island fault bisects Saddle Island and dips 30° to the northwest (Choukroune and Smith, 1985). More likely, however, the LVVSZ and the Lake Mead fault system are the major structures that control normal faults and detachments (Campagna and Aydin, 1994; Rowley, 1998). An alternative view is that of Anderson and others (1994) who suggested that the LVVSZ bounds the northern side of westward-flowing ductile rocks.

Previous efforts to determine the location of the LVVSZ beneath the alluvial deposits of Las Vegas Valley were limited to gravity studies (Plume, 1989; Campagna and Aydin, 1994). Longwell and others (1965) indicate the LVVSZ as a N. 60° W.-striking fault extending northeast from Frenchman Mountain to Indian Springs, Nevada. Campagna and Aydin (1994)

showed the LVVSZ as several parallel strands stepping across Las Vegas Valley to a point south of Frenchman Mountain and cutting across the northern tip of the McCullough Mountains. Our interpretation of the LVVSZ indicates two non-parallel strike-slip faults creating two sets of pull-apart basins beneath Las Vegas Valley (figs. 3*A* and 4; Langenheim and others, chapter B this report). The northern strand of the LVVSZ as shown by Campagna and Aydin (1994) coincides approximately with our interpreted northern strand of the LVVSZ. We, however, disagree with the location and orientation of the southernmost strand they proposed, shown cutting across the northern McCullough Mountains. Wallin and Duebedorfer (1995) reported no noteworthy fault in that area. Furthermore, the aeromagnetic anomalies generated by the volcanic rocks in the northern McCullough Mountains do not show evidence of strike-slip faults in this area (map C; fig. 5). Instead, we propose that a strand of the LVVSZ cuts across the valley and is aligned with the Frenchman Mountain fault segment that bounds the southern margin of Frenchman Mountain (fig. 3*A*; map A). The two strands of the LVVSZ could be contemporaneous; alternatively, they may be of different ages. The left-lateral strike-slip faults of the Lake Mead fault system apparently exhibited similar behaviors: older strands strike more easterly (N. 60°–70° E.) than do the younger faults. This relationship may also apply to strands of the LVVSZ, where older strands may be characterized by more westerly trends.

State Line and West Spring Mountains Fault Zones

The State Line fault zone (Blakely and others, 1998) is a northwest-striking, right-lateral, strike-slip fault zone that extends through Pahrump Valley (fig. 3*A*; map A). The fault has also been referred to as the State Line fault (Hewett, 1956), Pahrump Valley fault zone (Wright and others, 1981; Hoffard, 1991), Amargosa River fault zone (Donovan, 1991), and the Amargosa fault zone (Schweickert and Lahren, 1997). The fault zone has a topographic expression that extends about 175 km northwest from Ivanpah Valley, through Mesquite Valley and Ash Meadows, and may continue northwestward to define the northeast flank of the Funeral Mountains. The fault zone is structurally complex, and based on gravity data, the bedrock basement concealed beneath alluvial valleys consists of a series of narrow, steep-sided sub-basins, similar to Las Vegas Valley. The sub-basins have been interpreted as transtensional or pull-apart basins, caused by en-echelon right steps during strike-slip faulting (Blakely and others, 1998).

In the Pahrump Valley, the bedrock basement surface is relatively flat along the valley margins, but within the central part of the valley, two northwest-trending steep-sided basins are separated by a basement gravity ridge (Blakely and others, 1998; figs. 6*B* and 9*B*). Based on gravity data, the southwestern basin is 2 km deep, and the northeastern basin is about 5 km deep. The basement gravity ridge is exposed on the surface

in the Pahrump area as a gravel-capped ridge that extends northwest through the Mound Spring quadrangle (Lundstrom and others, 2003) in the southwestern part of the quadrangle. The ridge is interpreted as a horst block, or as a compressional structure caught up between reverse faults along this segment of the fault zone. The basement ridge may be an important feature controlling ground-water flow in Pahrump Valley. Blakely and others (1998) suggested that faults bounding the northeast margin of the ridge may act as a ground-water barrier, where ground water flowing from the north and east is brought to the surface. This is based on the distribution of springs and paleospring discharge deposits restricted to areas along the northeast side of the ridge.

The West Spring Mountains fault zone (fig. 3*A*; map A) (Anderson and others, 1995b) is the range front fault zone along the southwestern Spring Mountains. The overall fault zone has a sinuous trace, but most strands strike north to northwest. The southern end of the fault zone merges with the State Line fault zone in Pahrump Valley. The fault was determined to be mostly dip-slip, and it accommodated west-side-down movement. The fault zone cuts Quaternary deposits as young as late Pleistocene and has a strong geomorphic expression.

Cenozoic Landslide Deposits

Well-consolidated Cenozoic landslide deposits are on the flanks of the Spring Mountains. These deposits are frequently referred to as gravity-slide blocks, because most formed by downslope movement influenced by gravity. The landslide deposits are bound at their base by low-angle shear-slip surfaces, denoted on map A by a single-hachured fault, with the hachures on the upper plate. Landslide deposits are generally of two textural types in the quadrangle. The first type consists of kilometer-size blocks of bedrock units that have maintained their stratigraphic order during landsliding, except for local breccia zones along their slip surfaces. The second type is composed predominantly of megabreccia. Although megabreccia texture is commonly associated with large landslide block deposits and large blocks are locally observed within megabreccia deposits, the landslide deposits discussed below were categorized based on their dominant textural type.

An example of the first type of deposit is a large, composite landslide flanking the southwest Spring Mountains near the mouth of Trout Canyon (Burchfiel and others, 1974) (map A). In this area, large landslide blocks (as much as 4 by 6 km in diameter), consisting mostly of the Mississippian to Permian Bird Spring Formation, slid over Cambrian rocks in the upper plate of the Lee Canyon thrust, and one of these landslide blocks partly conceals the southern segment of the thrust. Burchfiel and others (1974) suggested that the stratigraphic coherency of these large blocks indicated they moved downslope somewhat slowly.

Another example of this type was recently recognized (W.R. Page, new mapping, this report) in the Red Rock Wash

area, at the northeast edge of Blue Diamond Hill in the south-central part of the map (map A). This feature is characterized by a 5 by 6 km block of Kaibab and Toroweap Formations, just southeast of the Red Rock Canyon visitor's center. We interpret this block to have detached and slid eastward from Blue Diamond Hill due to faulting and local relief along the La Madre fault scarp. The basal shear-slip surface of the landslide block is bowl-shaped, and rocks in the upper plate dip to the northeast and are highly deformed by closely spaced, high-angle normal faults (map A) that most likely sole into the basal shear plane. In contrast, rocks in the stable block of Blue Diamond Hill dip gently to the west and are relatively unde-formed. A north-trending, east-verging overturned anticline at the eastern margin of the landslide block probably accomo-dated local compression at the toe of the landslide.

Lone Mountain, an isolated 1-kilometer-diameter block of Devonian and Mississippian rocks east of La Madre Mountain in the central part of the map, is also interpreted as a landslide deposit of this type. Geophysical studies indicate that Lone Mountain is underlain by low-resistive, low-den-sity rocks (Zohdy and others, 1992; Langenheim and others, chapter B, this report). Zohdy and others (1992) interpreted these rocks as the Aztec Sandstone, saturated with water, and possibly the Chinle or Moenkopi Formations. Alternatively, these rocks may be Tertiary and Quaternary basin-fill deposits. These data suggest that Lone Mountain is not rooted by older Paleozoic rocks and is a landslide block that detached from the La Madre Mountain area and slid over either Mesozoic units or Tertiary and Quaternary basin-fill deposits during Neogene extension.

West of Willow Creek and east of the Nye–Clark County line in the northwestern part of the quadrangle, W.R. Page (map A) mapped several low-angle-normal faults in Ordovi-cian through Mississippian rocks that may be part of a large composite landslide complex, shed off the northwest Spring Mountains. Along one low-angle fault, a 2-kilometer-diameter block composed of coherent upper-plate Devonian and Mis-sissippian rocks overlies Ordovician rocks in the lower plate (map A). Along other low-angle faults in this area, Mississip-pian rocks in the upper plate are juxtaposed against Devonian rocks in the lower plate. These low-angle faults may represent shear-surfaces of a much larger composite landslide deposit that extends westward into the adjacent Death Valley Junction 1:100,000-scale map, where similar low-angle normal faults are present. One of these faults is the Point of Rocks detach-ment fault (Abolins, 1999). K-Ar ages reported for volcanic ash beds interbedded with synextensional basin deposits in the upper plate of the fault range from 13 to 10 Ma, and these ages date, or slightly post-date, movement on the fault (Abolins, 1999). Abolins (1999) suggested upper plate rocks were prob-ably derived from the northwest Spring Mountains during extensional unroofing of the range.

The second type of landslide deposit is chaotically mixed megabreccia consisting of clasts of bedrock units in a com-minuted matrix of the same material. Internal texture of the megabreccia landslides suggests that they formed as rock

avalanche deposits and were emplaced rapidly downslope. Page and others (1998) described megabreccia deposits along the eastern Spring Mountains that they called the Blue Dia-mond landslide. The megabreccia is predominantly derived from Paleozoic rocks interpreted to have slid nearly 10 km from their proposed source area above the Wilson Cliffs. Although the majority of the Blue Diamond landslide deposit is labeled on map A as unit **MDu** (Sultan Formation), other rock units are identifiable in the deposit including rocks of the Monte Cristo Group. Axen (1985) mapped similar megabrec-cia deposits in the La Madre Mountain area; these may be part of the Blue Diamond landslide.

Other megabreccia landslide deposits are exposed on the northwest flank of the Spring Mountains, between Willow and Cold creeks. The upper plate of this landslide consists of megabreccia deposits derived mostly from the Cambrian Bonanza King Formation which slid northeastward above Ordovician rocks. We interpret that these deposits were shed off a splay of the La Madre fault (map A).

The age of emplacement for landslides on the flanks of the Spring Mountains is poorly constrained, but we agree with Burchfiel and others (1974) and Axen (1985) that most of these well-consolidated landslides are Cenozoic in age and were emplaced during maximum episodes of extension in the Miocene (Abolins, 1999; Page and others, 1998). Miocene landslide deposits are reported in areas adjacent to the Las Vegas quadrangle and in other parts of the southern Basin and Range Province. Guth and others (1988) reported megabreccia landslide deposits on the west side of the Sheep Range that were emplaced between 17 and 12 Ma. Landslide deposits on the flanks of the northern Mormon Mountains, about 100 km northeast of the quadrangle, are 12.5–15 Ma (R.E. Anderson, U.S. Geological Survey, written commun., 1998). Minor and Fleck (1994) reported landslide deposits near Beatty, Nevada, 100 km to the west of the quadrangle, that were emplaced between 11 and 12 Ma. Krieger (1977) documented landslide deposits in the basin and range of southeastern Arizona that are encased in Miocene basin-fill sediments. The occurrence of Miocene landslide deposits throughout the southern basin and range region suggests they shared a common genesis, and were generated in response to high relief formed by fundamental processes of crustal extension. These processes include development of basin-range topography through high- and low-angle normal and strike-slip faulting and development of positive areas related to the rise of plutons and core complexes.

Quaternary Faults

Introduction

Quaternary fault activity in the Las Vegas quadrangle is indicated by numerous faults that displace Quaternary deposits at the surface. A wide range of studies on Quaternary faults in the Las Vegas region were presented at a seismic hazards

conference in 1996 (dePolo, 1998). The overview and discussion below emphasizes mapped Quaternary faults in the Las Vegas and Pahrump Valleys, which are the two basin areas dominated by Quaternary surficial deposits. Because Quaternary faulting can only be recognized where Quaternary deposits and surfaces exist in appropriate positions, it is possible that unrecognized Quaternary faulting could be associated with some of the bedrock structures noted above, especially the high angle normal faults.

Las Vegas Valley

A prominent set of fault scarps exist in the central part of the Las Vegas Valley (fig. 3*A*). These intra-valley faults, including the Decatur, Valley View, and Cashman faults (Slemmons, 1992), were first systematically mapped by Bell (1981) who studied their relationships to subsidence due to ground-water withdrawal and to tectonics. At the latitude range of central Las Vegas, the faults are generally north-south trending and form large (ranging in height from 12–55 m) scarps in fine-grained sediments. The scarps and sediments are now mostly covered by urbanization, but the fault scarps are still obvious in the urban landscape. The basic geologic relations of these faults were reported in the four Las Vegas 7.5' quadrangles (Bingler, 1977; Matti and Bachuber, 1985; Matti and others, 1987, 1993) through analysis of aerial photographs taken before 1950.

The north-south trend of these intra-valley faults changes to southeastward in the southern part of the Las Vegas Valley and to northeastward in the northern part of the valley. In the southeast corner of the quadrangle, the Whitney Mesa fault zone (fig. 3*A*) has several strands forming a composite scarp 55 m high just south of Whitney Mesa, but this fault system is apparently buried by the caliche-cemented gravels (**Qao**) which cap Whitney Mesa. However, discontinuous fractures aligned on this fault zone cut the caliche and are locally filled with travertine where sample WM1A-A and WM1A-B yielded U-series dates of 363 ± 30 ka and 317 ± 19 ka (table 2).

The north end of the Decatur fault curves northeast to become the Eglington fault (fig. 3*A*). Similarly, the north ends of the Valley View and Cashman faults turn to northeasterly trends. Thus, the wrap-around pattern of the intra-valley faults is subparallel to both the west margin of the Frenchman Mountain block and its subsurface gravity signature, as well as to the concave eastward curvature of the Keystone thrust. Whether there is any genetic reason for this coincidence is unknown and deserves further study.

All of the above intra-valley faults are in fine-grained deposits inferred to at least partly include deposits associated with past ground-water discharge that were documented as such north of the Eglington fault (Haynes, 1967; Quade, 1986). The faults form discontinuities and barriers to the stratigraphically controlled basinward ground-water flow and thus may have been significant in localizing past ground-water discharge. A modern analogue is the area of historically active

artesian springs of the Las Vegas Valley which are along the Valley View fault.

Conversely, the faults and associated fine-grained deposits are known to control the patterns of subsidence related to ground-water overdrafts in the Las Vegas Valley (Bell, 1981; Amelung and others, 1999). Mifflin (1998) attributes most movement on these intra-valley faults to differential compaction due to dewatering of compressible fine-grained sediments to the east of each of the above faults. Bell and dePolo (1998) showed that the magnitude of apparent offset from scarp height cannot be explained by differential compaction alone; therefore a significant tectonic component is likely. It is noteworthy that a large area of recent historic subsidence measured by Amelung and others (1999) is on the northwest, upthrown side of the Eglington fault, which bounds most of the subsidence. This is opposite the sense needed to explain down-to-southeast fault movement by prehistoric ground-water compaction.

The Eglington fault forms a broad low-angle scarp about 12 m high that steepens downslope toward its base. The surface expression of this fault scarp coincides with a gravity ridge bounding two gravity lows (chapter B) and with offsets in the bedrock alluvial contact and in the valley-fill alluvium at depth (Plume, 1989). The Eglington fault may be bounded at its northeast end by the Las Vegas Valley shear zone (LVVSZ).

The LVVSZ is delineated in the subsurface by a steep gravity gradient (discussed in chapter B), but it does not have a clear surface fault expression. However, the Corn Creek Springs fault (Haynes, 1967; Bell and others, 1999) is aligned on the LVVSZ and has surface expression as it cuts associated fine-grained deposits associated with past ground-water discharge (Quade and others, 1998). Parallel to this zone, and to the northeast, the northwest-trending frontal fault of the Sheep Range also shows Quaternary faulting that juxtaposes bedrock against Quaternary fan gravels.

The Frenchman Mountain fault (Anderson and O'Connell, 1993) (fig. 3*A*) is at the western range front of Frenchman Mountain, but in detail has various non-aligned short segments cutting various ages of alluvial fans (Peck, 1998) ranging in age from middle Pleistocene to early Holocene (map A). The steep gravity gradient west of the range front fault (discussed in chapter B) coincides with the entire Frenchman Mountain piedmont and indicates the main range-bounding fault is 1–2 km west of the range front. This is consistent with the occurrence of outcrops of the Muddy Creek(?) Formation which are overlain on this piedmont by thin Quaternary pediment deposits.

To the west of the Las Vegas Valley, in the area where projections of the La Madre, Turtlehead Mountain, and Brownstone Basin faults and Bird Spring thrust all converge on Red Rock Wash, there is a large (about 30 m high) east-facing fault scarp bounding map unit **QTa**. Just north of the projection of this scarp, a late Pleistocene terrace of **Qai** has a gentle monocline that displaces its grade about 15 m down to the east. At the hinges of the monocline, there are fractures

and minor displacement probably related to the folding of the carbonate-cemented horizons of the surficial soil of **Qai**.

Pahrump Valley

Gravity investigations (discussed in chapter B) show an elongate northwest-trending horst along the Nevada-California boundary, flanked by northwest-trending basins on both sides. The horst and flanking basins, and their approximate alignment with similar subsurface features beneath the Amargosa Desert, were interpreted by Blakely and others (1999) as transpressional and pull-apart features, respectively, along the proposed, northwest-trending right-lateral State Line fault system.

The Quaternary geology is consistent with this structural interpretation. The gravity high coincides with the surface occurrence of **QTa**, which forms a northwest-aligned chain of hills. This alignment is traversed and cut by the southwestern-sloping, modern and late Quaternary drainage system tributary to the playa (**Qpy**) in the southwest corner of the quadrangle. A large, complex, southwest-facing, northwest-trending escarpment as much as 20 m high coincides with the outcrop of **Qbu**. The scarp is about 2 km northeast of and sub-parallel to the hills composed of unit **QTa** that are along the crest of the gravity ridge. Though Hoffard (1991) and Anderson and others (1995a) considered this scarp as a fault scarp of a down-to-the-southwest normal fault, we interpret an opposite, northeast-down sense of displacement based on subsurface geology indicated by the gravity gradient. Moreover, in gully exposures near Brown Spring and in the adjoining Hidden Hills Ranch quadrangle, we can trace continuous unfaulted beds of **QTbo** from gullies cut into the scarp to a few hundred meters southwest of the scarp. The beds exposed in the scarp area dip 5°–15° northeast into the scarp, and the main scarp is held up by the resistant limestone bed of **Qbw**. Northeast dips increase as much as 20° as the scarp is crossed to the southwest, suggesting the scarp may be localized along the east side of a northeastward-facing monocline. The northeast dips on middle Pleistocene strata in a direction opposite the southwest topographic slope are consistent with middle to late Quaternary uplift in the area of the horst. Because surface fault scarps do not cut the predominantly Holocene sediments in the valley between the scarp and the hills composed of unit **QTa**, we infer a buried, intermittently active fault with down-to-the-northeast displacement under the intervening sediments, consistent with both the gravity data and surface geology.

In the area of fine-grained sediments between the playa and the State Line horst in the southwest corner of the map, there are numerous curving lineaments that are prominent on aerial photographs (Lundstrom and others, 2003). We interpret these lineaments to be tension cracks accentuated by vegetation; they lack surface displacement and are difficult to identify on the ground. They may be related to some combination of dessication, evaporation, and ground-water discharge through the playa surface, and (or) extension related to the hypothesized strike-slip and pull-apart tectonics.

About 9 km east of the Pahrump Valley playa and apparently splaying off the State Line horst is a north-northwest-striking system of left-stepping en-echelon west-facing scarps which cut both older and younger fine-grained deposits, as well as **Qai** fan gravel. As in Las Vegas Valley, the faults probably formed barriers to valleyward ground-water flow that favored past ground-water discharge with which the fine-grained deposits are associated. Similarly, a fault bounds and probably controlled past discharge associated with fine-grained deposits forming badlands along the highway near Pahrump (dePolo and others, 1999). Cracks and related piping features were observed and mapped along the upper parts of scarps in the late Pleistocene ground-water discharge deposits. These are probably a result of subsidence from historic ground-water withdrawal similar to that documented in the Las Vegas Valley (Bell, 1981).

About 5–8 km to the east of the State Line horst is the south end of the West Spring Mountain fault zone (fig. 3*A*; Anderson and others, 1995a), which forms a prominent scarp and graben in **Qai** fan gravels. The west-facing scarp forms a sharp 90° bend about 1 km from the range front. Further to the northwest, more west-facing scarps form left-stepping segments several kilometers from the range front.

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Chapter B
Geophysical Framework of the Las Vegas 30′ × 60′ Quadrangle

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Introduction

An important objective of geologic mapping is to understand how surficial structures and stratigraphy project into the subsurface. Geophysical data and analyses are useful tools for reaching this objective. The gravity and aeromagnetic maps provide a three-dimensional perspective to the geologic map of the Las Vegas 30′ × 60′ quadrangle. The isostatic gravity anomaly map (map B, map sheet 2) reflects density variations in the upper and middle crust. Interpretation of these anomalies provides information on the depth of Cenozoic basins and on the nature of pre-Cenozoic basement rocks. The aeromagnetic map (map C, map sheet 2), in contrast, is more sensitive to the distribution of magnetic rocks, primarily those containing magnetite. In the Las Vegas quadrangle, these rocks are Tertiary igneous and Precambrian crystalline rocks. In particular, aeromagnetic anomalies mark abrupt spatial contrasts in magnetization that can be attributed to lithologic boundaries, potentially caused by faulting of these rocks. These two geophysical maps, in concert with information from geologic mapping, wells, and other geophysical data, provide constraints on the subsurface geology.

Aeromagnetic Data

The aeromagnetic map (map C) is based on measurements from east-west flight lines acquired in two separate surveys (U.S. Geological Survey, 1979; 1983). Both surveys were flown at 300 m (1,000 ft) above the ground along flight lines spaced 1.6 km (1 mi) apart. The data were adjusted to a common datum and then merged by smooth interpolation across a buffer zone along the survey boundaries. Saltus and Ponce (1988) published these data as a mosaic at a scale of 1:250,000.

To help delineate trends and gradients in the aeromagnetic data, we calculated magnetization boundaries in the following way: first, in order to emphasize the edges of shallow magnetic sources, we subtracted a numerically derived regional field from the actual data. The regional field was computed by analytically continuing the aeromagnetic data to a surface 100 m higher than actually measured, an operation that tends to smooth the data by attenuating short-wavelength anomalies (Blakely, 1995). The resulting residual aeromagnetic field accentuates those anomalies caused by shallow sources and is shown in figure 5. Second, the resulting residual aeromagnetic field was mathematically transformed into pseudogravity anomalies (Baranov, 1957); this procedure effectively converts the magnetic field to the equivalent "gravity" field that would be produced if all magnetic material were replaced by proportionately dense material. Third, the horizontal gradient of the pseudogravity field was calculated everywhere by numerical differentiation. Lastly, locations of the locally steepest horizontal gradient were determined numerically (Blakely and Simpson, 1986). These locations occur approximately over vertical or near-vertical contacts that separate rocks of contrasting magnetic properties.

Isostatic Gravity Data

The isostatic gravity map (map B) was created from more than 2,500 gravity stations. The details of the data acquisition, sources, and processing can be found in Langenheim and others (1999). In general, gravity highs occur over exposed or shallow pre-Cenozoic rocks, whereas gravity lows occur over thick sections of Cenozoic sedimentary and, in some cases, volcanic rocks. The major features on the isostatic gravity map are the gravity lows associated with the Las Vegas and Pahrump Valleys and the gravity high over the Spring Mountains. The isostatic gravity values range from a low of -33 mGal (centered over the Gass Peak SW 7.5′ quadrangle) to a high of +25 mGal over the highest part of the Spring Mountains near Charleston Peak. The gravity data provide information on major structural features such as the Las Vegas Valley shear zone (LVVSZ), the State Line fault zone, and the Frenchman Mountain fault (fig. 4).
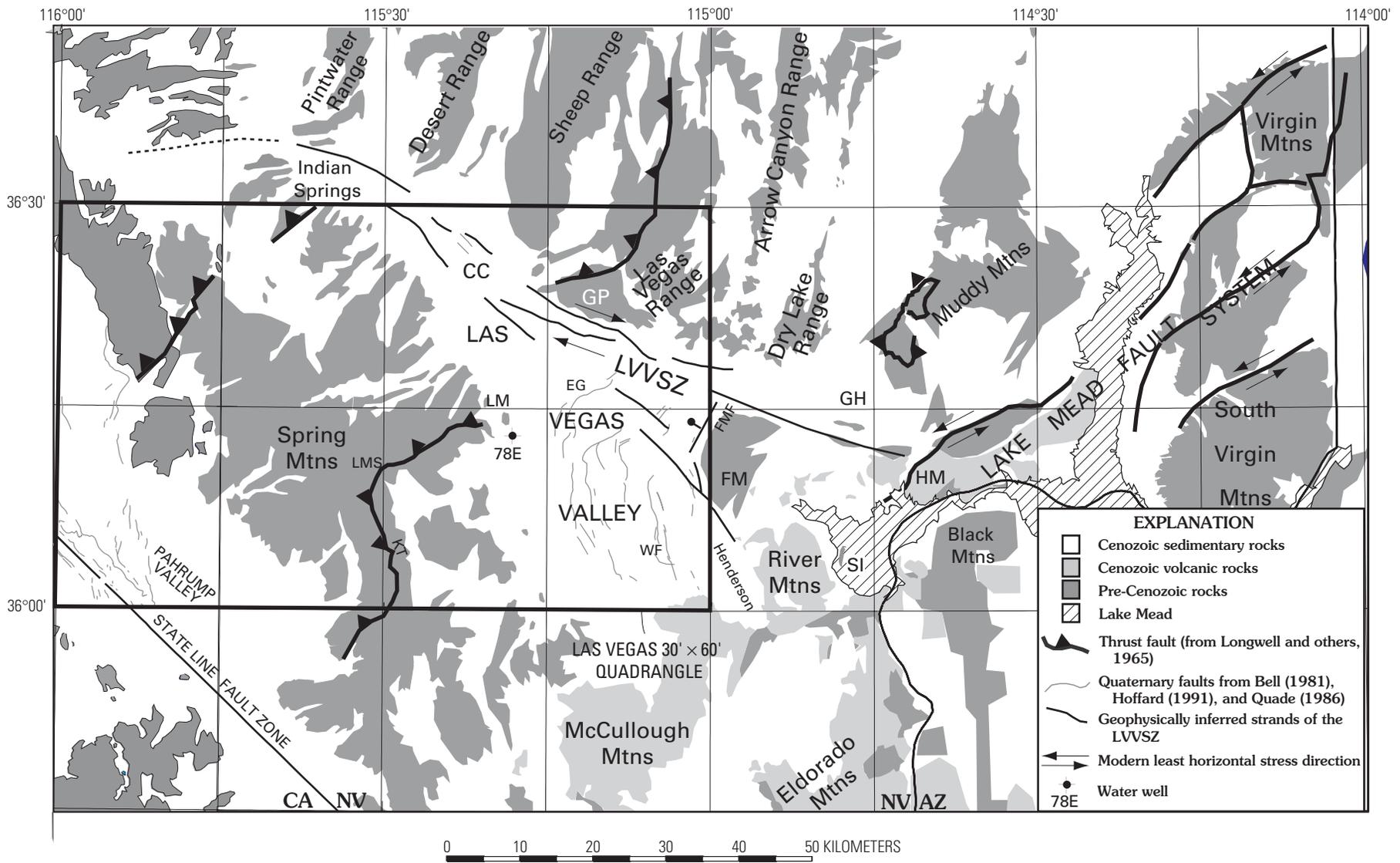
Figure 4. Index map of the Las Vegas region showing simplified geology (modified from Longwell and others, 1965) and major structural features. CC, Corn Creek Springs; EG, Eglington fault; FM, Frenchman Mountain; FMF, Frenchman Mountain fault; GH, Gale Hills; GP, Gass Peak; HM, Hamblin Mountain; KT, Keystone thrust; LM, Lone Mountain; LMS, La Madre Spring; SI, Saddle Island; WF, Whitney Mesa fault. Heavy black box outlines the Las Vegas 30′ × 60′ quadrangle.

The method of Jachens and Moring (1990) was used to separate that part of the gravity field caused by Cenozoic deposits from that caused by variations in the density of the basement rocks (defined here as pre-Cenozoic). The method was modified to include constraints based on well information and depths to basement derived from other geophysical data (B.A. Chuchel, written communication, 1995). Much of the following discussion is based on the results of the inversion and summarizes various U.S. Geological Survey Open-File reports (Blakely and others, 1998; Langenheim and Jachens, 1996; Langenheim and others, 1997; Langenheim and others, 1998).

Basins

Las Vegas Valley

The broad alluvial valley of Las Vegas is characterized by a roughly rectangular, northwest-trending gravity low centered near the northern part of the valley. Another low of substantially less magnitude (minimum gravity value -15 mGal) lies to the southeast within the Las Vegas SE 7.5' quadrangle (southeast corner of the map). The southwestern part of the valley is marked by higher isostatic gravity values (exceeding 0 mGal), suggesting that this part of the valley is underlain by a shelf of relatively shallow pre-Cenozoic basement (Langenheim and Jachens, 1996). This region also corresponds to a broad aeromagnetic high, suggesting the presence of magnetic Precambrian basement rocks at depths of 3–4 km. To the east, aeromagnetic anomalies indicate shallower sources, probably buried Tertiary volcanic rocks, similar to those exposed in the McCullough Mountains.

The gravity field over Las Vegas Valley was inverted for basin thickness using a density-depth function derived from borehole porosity data (table 3), available well data that constrain the thickness of Cenozoic fill, and depths to basement from seismic-reflection profile LV-4 (Langenheim and others, 1998). The resulting distribution of basin sediments (fig. 6*A*), in general, mirrors the gravity field, although the deepest part of the basin is not over the lowest gravity values in the basin, but 5 km west of Frenchman Mountain. The apparent mismatch is caused by allowing for variations in basement density during the inversion method. Density measurements from hand samples indicate that significant density variations are possible between various basement rock types (table 4).

Plume (1989) also produced a basin thickness map for Las Vegas Valley that differs from that shown in figure 6*A*. Our maximum basin thickness is more than 4 km, much deeper than his maximum estimated depth of 1.5 km. Our model is based on a more detailed gravity data set, independent depths to basement, and densities that increase with depth. Plume's map is based on a single density contrast between basement and basin fill (-400 kg/m^{3}).

The basement surface under Las Vegas Valley is complex and consists of elongated sub-basins with N. 70° W., N. 50° W., and N. 40° E. trends (fig. 6*A*). Many Quaternary fault scarps mapped within the valley fill coincide with changes in basin thickness (for example, the Eglington fault, EG on fig. 6*A*) and edges of residual aeromagnetic anomalies (fig. 5). Cross section *D–D′* (fig. 9*A*) illustrates the correlation of mapped faults with steps in the basement surface, especially

Table 3. Density-depth functions used to determine thickness of basin-fill deposits.

Depth range (meters)	Density contrast above basement rocks (kg/m3)			
	Las Vegas Valley		Pahrump Valley	
	sedimentary	volcanic	sedimentary	volcanic
0-200	-600	-450	-650	-450
200-600	-450	-400	-550	-400
600-1,200	-250	-250	-350	-350
>1,200	-250	-250	-250	-250

Table 4. Densities (kg/m³) and magnetic susceptibilities (10^{-3} SI units) of basement rocks.

No. samples	Density range	Average density	Susceptibility range	Average susceptibility
Precambrian crystalline rocks				
3	2,630–2,940	2,830	0.25–2.85	13.1
Paleozoic sedimentary rocks				
33	2,240–2,850	2,680	0.00–0.00	0.00
Mesozoic sedimentary rocks				
13	2,210–2,670	2,500	0.00–0.01	0.00

those faults that strike perpendicular to *D–D'*. The curved geometry of the faults suggests that the LVVSZ may have played a role in their formation (Langenheim and others, 2001). The origin of these faults is controversial; they may be tectonic or caused by differential compaction of sediments (due to ground-water withdrawal) or both (Bell, 1981). Interestingly, recent subsidence as measured by interferometric synthetic aperture radar (InSAR) data, has occurred on the upthrown side of the Eglington fault (Amelung and others, 1999), opposite to the main sense of offset measured by seismic data (Plume, 1989) and inferred from gravity data (this study).

The deepest part of the basin under Las Vegas Valley trends northeast, parallel to the Frenchman Mountain fault system (Bell, 1981). This fault system is characterized by multiple shear planes (both along the bedrock-valley interface and within the alluvium) that (1) range in dip from as low as 35° to near vertical and (2) young westward (Peck, 1998). The Frenchman Mountain fault scarps (west of the faults mapped along the bedrock contact) coincide with the strong gravity gradient. The inversion suggests that the main basin-bounding fault is steep, with an average dip of 60°–65° E, and it is 1–2 km west of the bedrock contact (see also *D–D'*, fig. 9*A*). The gravity gradient also coincides with a strong aeromagnetic gradient, which suggests that the fault also offsets magnetic Precambrian crystalline rocks.

The northern margin of Las Vegas Valley is marked by a strong gravity gradient. A weak magnetization boundary coincides with the location of the steep gravity gradient south of Gass Peak (fig. 7). However, caution should be exercised in interpreting this boundary because the east-west survey flightlines cross the LVVSZ at a highly oblique angle, and the filtering process may have produced artifacts in the residual data. Plume (1989) suggested that the gravity gradient south of Gass Peak marks the buried trace of the LVVSZ. Campagna and Aydin (1994) argued that the northern margin marks one of three strands of the shear zone, on the basis of their analysis of the gravity field. Langenheim and others (1998, 2001) concurred that the steep gradient south of Gass Peak (GP) represents the LVVSZ, and indicates at least two closely spaced strands of the shear zone along the northern margin of the basin. They infer another strand cutting across the valley to the south and connecting with a fault mapped along the southwest margin of Frenchman Mountain (figs. 4 and 6*A*).

Corn Creek Springs

In contrast to the basin morphology beneath Las Vegas Valley, the basin northwest of Corn Creek Springs is very narrow and elongate, no more than 5 km wide, and characterized by sub-parallel, N. 50° W.-trending edges (fig. 4). The basin is as much as 3 km deep. Corn Creek Springs itself is located along the northeastern gravity gradient of the Corn Creek Springs gravity low. The gravity gradient coincides approximately with the Corn Creek Springs fault(s) which documents

late Quaternary movement (Haynes, 1967; Quade, 1986; Bell and others, 1999). An electrical resistivity cross-section derived from inverted audio-magnetotelluric soundings shows a ridge of somewhat higher resistivities under Corn Creek Springs relative to those associated with the fine-grained Las Vegas Valley deposits (Pierce and Hoover, 1988). The two margins of the basin may be related to the LVVSZ. Campagna and Aydin (1994) interpreted the Corn Creek Springs gravity low as reflecting a pull-apart basin.

Pahrump Valley

Another prominent and compound gravity low lies on the western side of the Spring Mountains over Pahrump Valley. The low is bifurcated by a northwest-trending gravity high that reaches values as high as -16 mGal. The low to the northeast of the gravity ridge has values as low as -32 mGal; the low to the southwest, as low as -24 mGal. These anomalies are elongated in a northwestward direction. The ridge defined by gravity coincides with outcrops of the oldest Quaternary deposits in this part of Pahrump Valley (**QTa**); the outcrops form a northwest-aligned chain of hills. The northeastern margin of the gravity ridge roughly coincides with mapped Quaternary spring discharge (Lundstrom and others, 2003). Hoffard (1991) mapped numerous lineaments, some of which are probably faults, along the margins of the ridge and the two basins (fig. 6*B*), although she and Anderson and others (1995) infer an opposite sense of motion for the northeastern margin of the ridge than we do.

The basin thickness map for the Pahrump Valley (fig. 6*B*) is from Blakely and others (1998), who used inversion parameters somewhat different from those used east of the Spring Mountains (table 3). The resulting basin thickness map mirrors the gravity field. Most of the basement surface beneath Pahrump Valley is generally flat and lies within 200 m of the land surface. In the areas of the gravity low, however, the topography is punctuated by two prominent, laterally restricted basins (figs. 6*B* and 9*B*). The thickness of the Cenozoic deposits exceeds 2 km in the southwestern sub-basin and 5 km in the northeastern sub-basin. Some of the mapped faults and lineaments may result from fissuring of the sediments above the basement ridge (Lundstrom and others, 2003). These basins most likely formed as transtensional features along a right-lateral strike-slip fault system (State Line fault zone) paralleling the State line between California and Nevada (Wright, 1988; Blakely and others, 1998). The ridge between the basins may have formed because of slightly more westerly orientations of second-order faults within the State Line fault zone.

As discussed in this section, we prefer the interpretations for a compound gravity low and that the Pahrump Valley is underlain by two locally restricted sub-basins separated by a narrow, northwest-striking ridge of pre-Cenozoic basement. This model (fig. 6*B*) assumes that the density of basin fill is laterally uniform. It is likely, however, that the State Line fault zone has promoted ground-water flow, and this may have increased the densities of near-surface sedimentary deposits

along the fault zone by cementation and alteration. Thus, an alternative interpretation for the narrow gravity high invokes high-density sedimentary deposits along the State Line fault zone rather than basement topography. We favor the basement ridge interpretation, although the alternative of higher-density basin fill may contribute to the magnitude of the gravity high of the ridge.

Pre-Cenozoic Basement

The inversion method produces not only a basin thickness map but also a map of the basement gravity field (fig. 7). The basement gravity field reflects the density in pre-Cenozoic rocks exposed and buried beneath basins within the quadrangle. Caution should be used in interpreting basement gravity anomalies over basins without well control or independent control from other geophysical data. For this reason, the following discussion is limited to the ranges where pre-Cenozoic rocks are exposed or areas where several wells or geophysically determined depths to basement exist.

Spring Mountains

The Spring Mountains are characterized by a broad isostatic gravity high centered roughly over Charleston Peak, the highest mountain in southern Nevada. The highest densities measured in the Las Vegas quadrangle are from Precambrian gneiss, exposed on the western edge of Frenchman Mountain (lat. 36° 11′N., long. 115° 01′W.), and Paleozoic dolostones exposed in the Spring Mountains (Langenheim and others, 1998). No crystalline basement rocks are exposed in the Spring Mountains, but dolostones are part of the Paleozoic carbonate sequence exposed in the overall antidomal structure of the range. Two lines of evidence suggest that, although they undoubtedly contribute to the gravity high, these rocks are not the main source of the gravity high. First, the outcrop pattern of the carbonates does not coincide well with the gravity high. For example, the Nopah and Bonanza King Formations exposed in the upper plate of the Lee Canyon thrust are on the western flank of the gravity high, not over the gravity maximum. Second, the aeromagnetic map shows a broad positive anomaly over the Spring Mountains, suggesting that the Spring Mountains are underlain by magnetic rocks. Limestone and dolostone are not magnetic and therefore not a likely source of the aeromagnetic high. A more likely candidate for the large gravity and magnetic anomalies present over the mountain range is Precambrian metamorphic and (or) igneous rocks (Blank, 1987).

The basement gravity anomalies over the Spring Mountains generally do not coincide with mapped thrust faults, except for the area near La Madre Mountain. Here, a basement gravity low (relative to the Spring Mountains basement gravity high) trends northeast, approximately paralleling the Keystone thrust northeast of La Madre Spring. The low

appears to coincide spatially with outcrops of Aztec Sandstone in the lower plate. The Mesozoic sandstones are characterized by densities of 2,400 to 2,500 kg/m^3, which would produce a density contrast of -100 to -300 kg/m^3 with respect to the Paleozoic sedimentary sequence (table 4). Lone Mountain, composed of Devonian and Mississippian limestones, does not perturb the basement gravity low, suggesting that it is underlain by low-density rocks. This is supported by a profile of resistivity data across the southern end of Lone Mountain that indicates lower-resistivity rocks (< 150 ohm-m) at depths of 200–1,400 m (Zohdy and others, 1992). Zohdy and others (1992) suggested that the cause of the lower-resistivity rocks was Aztec Sandstone, saturated with water, and possibly silty Chinle and Moenkopi Formations. Another possible contributor to the gravity and resistivity low are Tertiary and Quaternary alluvial deposits.

Somewhat lower basement gravity values occur in the area of Blue Diamond. This area has the highest magnetic values (+660 nT, nearly 1,000 nT higher than the low in Las Vegas Valley) on the map. The intense, oblong aeromagnetic high is superimposed on the broader aeromagnetic high present over much of the Spring Mountains within the quadrangle. Our model (fig. 8) is consistent with unpublished data of H.R. Blank, Jr. (in Conrad and others, 1990), that indicates a strongly magnetic body at about 4 km depth. Our model indicates that the magnetic Precambrian basement extends to the middle crust. Conrad and others (1990) suggested that the broad anomaly was caused by elevated Precambrian basement ("core complex"). They also suggest that the superimposed anomaly in the area of Blue Diamond is caused by a quartz monzonitic body that post-dated thrusting and that intruded Paleozoic rocks, producing contact-metasomatic magnetite.

Frenchman Mountain

Only the westernmost part of Frenchman Mountain crops out in the Las Vegas 30′ by 60′ quadrangle. Values reach as high as 20 mGal, nearly as high as the Spring Mountains gravity high. Unlike the Spring Mountains, Precambrian crystalline rocks are exposed and readily explain the gravity high. The Precambrian rocks are also the probable source of the positive aeromagnetic anomaly. The aeromagnetic low associated with the Frenchman Mountain high wraps around the margin of Frenchman Mountain, suggesting that the source of the anomaly does not extend to great depths. Modeling of the aeromagnetic high indicates that the source rocks may extend to depths of 5 km beneath Frenchman Mountain (Langenheim and others, 1997).

Las Vegas Range

Although the same Paleozoic stratigraphy as that exposed in the Spring Mountains crops out in the Las Vegas and Sheep Ranges north of Las Vegas Valley, the gravity values are substantially lower than those in the Spring Mountains (+4 mGal
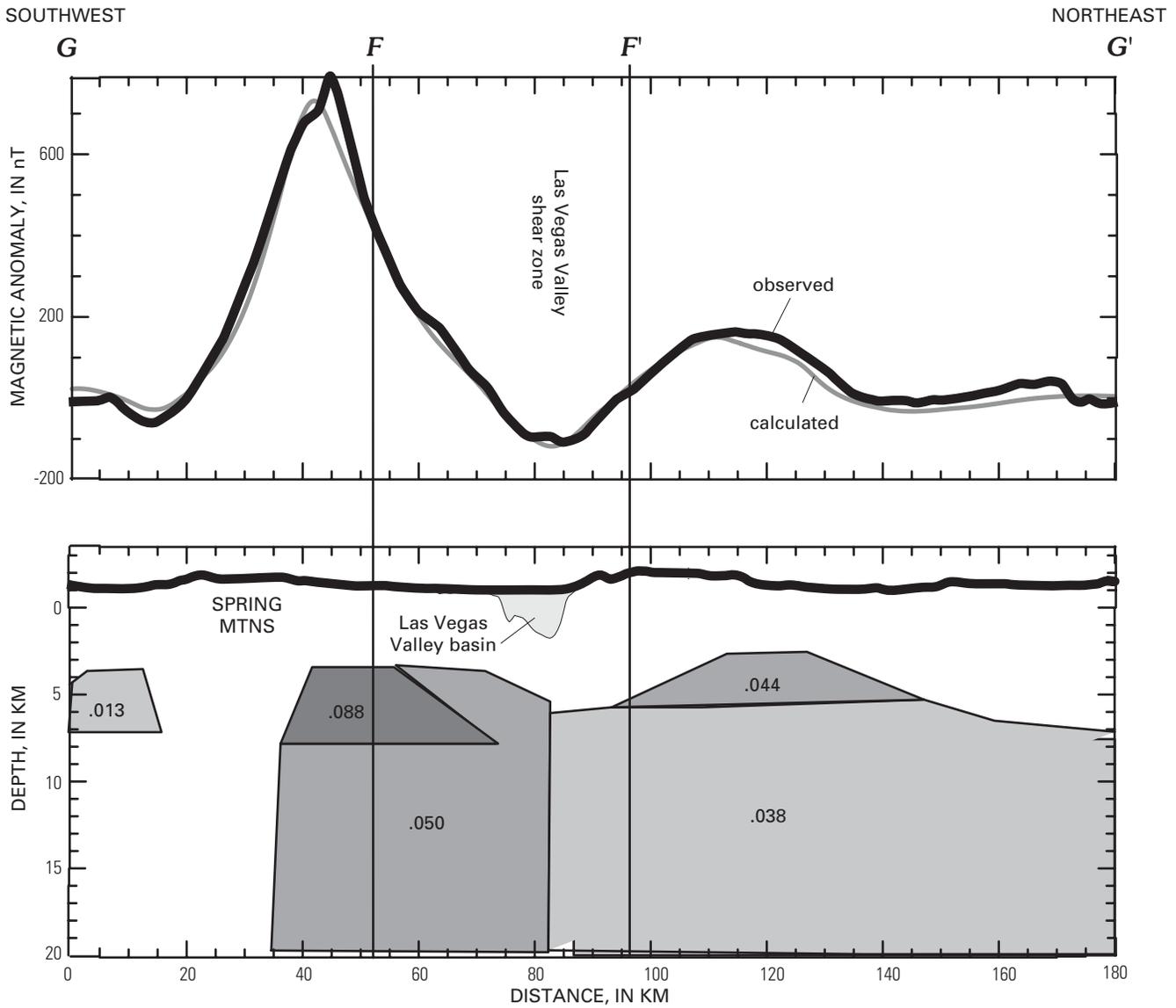
Figure 8. Magnetic model across Spring Mountains along extended cross section *F–F′* (*G–G′* marks the southern and northern boundaries, respectively, of the Las Vegas 30′ × 60′ quadrangle along the model). Numbers in model are magnetic susceptibilities in SI units. Vertical exaggeration 3×. Model assumes sources are greatly elongated in the east and west directions. Sediments in Las Vegas Valley basin are assumed to be non-magnetic. Maps B and C and figures 5, 6*A*, 6*B*, and 7 show locations of sections *F–F′* and *G–G′*.
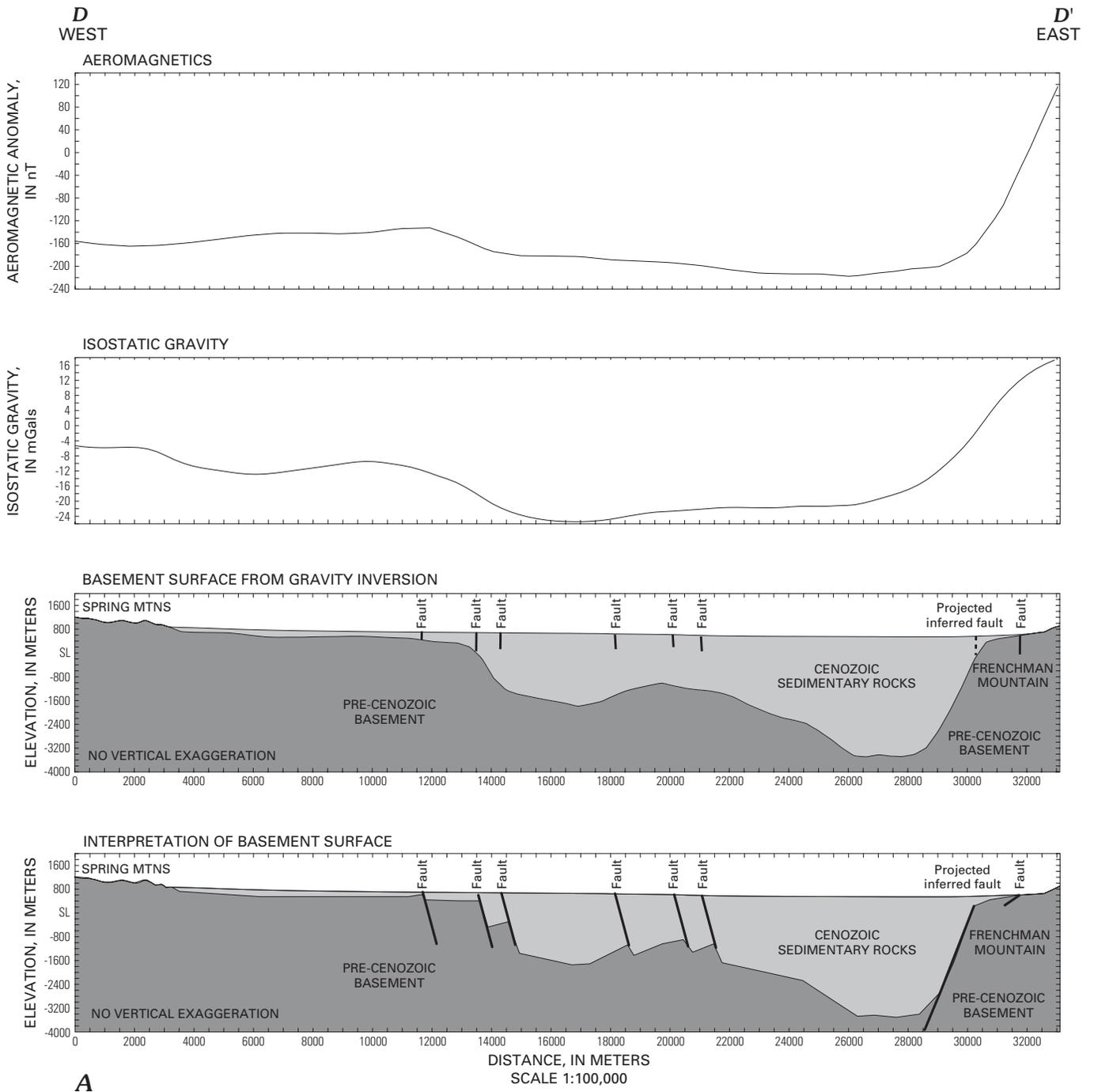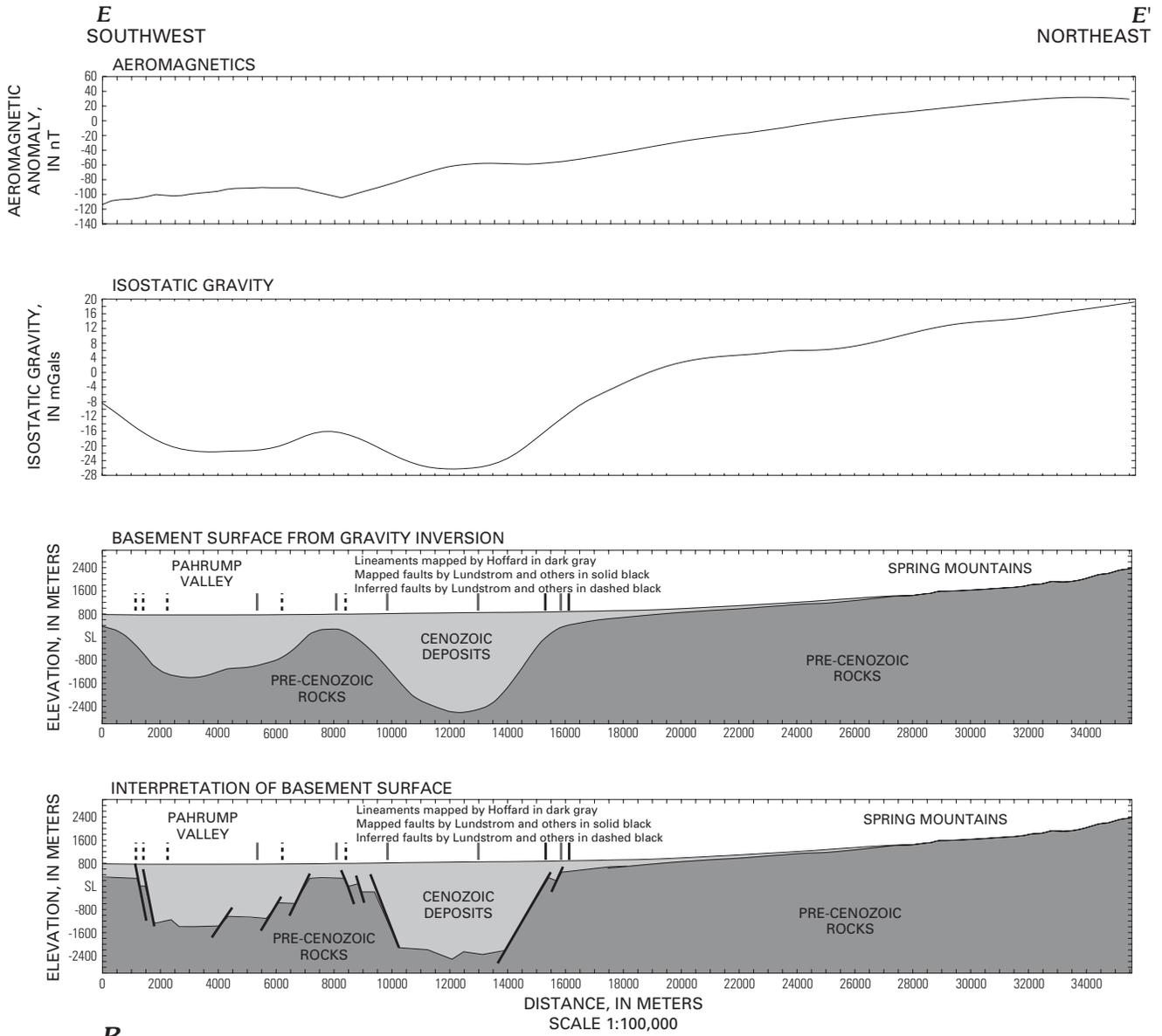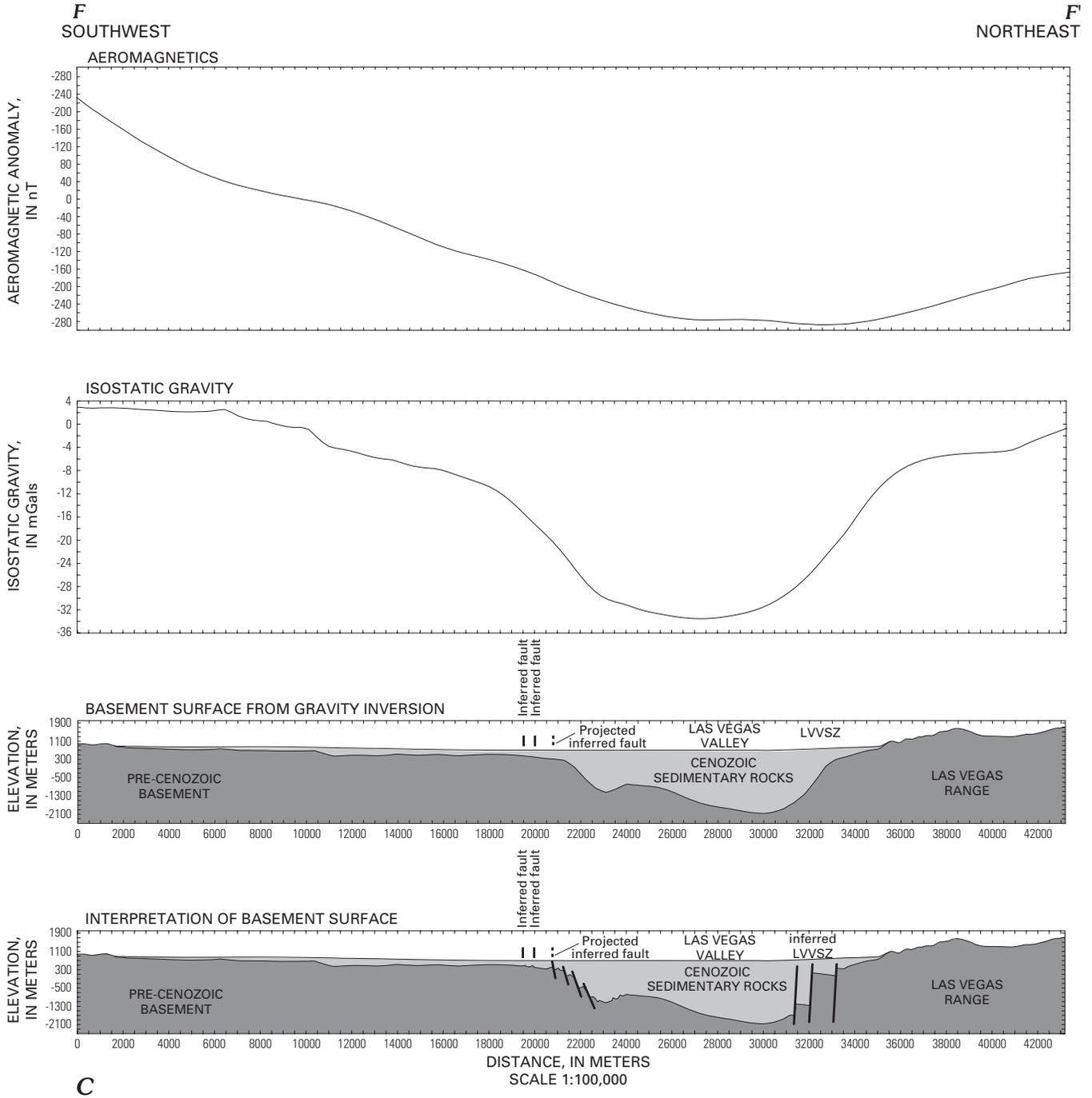
Figure 9 (above and following two pages).   *A*, Cross section *D–D′* extending east-west across Las Vegas Valley; location of cross section shown in Figure 6*A*. *B*, Cross section *E–E′* extending across Pahrump Valley; location of cross section line shown in figure 6*B*. *C*, Cross section *F–F′* extending northeast-southwest across Las Vegas Valley; location of cross section lines shown in figures 6*A* and 7. Maps B and C and figures 5, 6*A*, 6*B*, and 7 show locations of sections *D–D′*, *E–E′*, and *F–F′*.

E
SOUTHWEST

E'
NORTHEAST

AEROMAGNETICS

ISOSTATIC GRAVITY

BASEMENT SURFACE FROM GRAVITY INVERSION

PAHRUMP VALLEY

SPRING MOUNTAINS

Lineaments mapped by Hoffard in dark gray
Mapped faults by Lundstrom and others in solid black
Inferred faults by Lundstrom and others in dashed black

CENOZOIC DEPOSITS

PRE-CENOZOIC ROCKS

PRE-CENOZOIC ROCKS

INTERPRETATION OF BASEMENT SURFACE

PAHRUMP VALLEY

SPRING MOUNTAINS

Lineaments mapped by Hoffard in dark gray
Mapped faults by Lundstrom and others in solid black
Inferred faults by Lundstrom and others in dashed black

CENOZOIC DEPOSITS

PRE-CENOZOIC ROCKS

PRE-CENOZOIC ROCKS

DISTANCE, IN METERS
SCALE 1:100,000

B

C

versus +25 mGal), even allowing for the effects of the deep basin beneath Las Vegas Valley. Unlike the Spring Mountains, the Las Vegas Range is not characterized by a large aeromagnetic anomaly, suggesting that rocks composing the basement north of the Las Vegas Valley shear zone are different from the basement rocks beneath the Spring Mountains. Our model (fig. 8) suggests that the shear zone approximately marks a possible change in Precambrian basement type and may be a deep-seated feature.

Conclusions

Gravity and aeromagnetic data provide constraints on the nature of the subsurface in the Las Vegas 30′ × 60′ quadrangle. Important constraints from the gravity data include the basin configuration beneath Las Vegas and Pahrump Valleys. For both valleys, the alluvial deposits conceal a complex basement topography. Both basin configurations suggest that strike-slip faulting played an important role in basin formation. The large gravity and aeromagnetic highs over the Spring Mountains support the deep-rooted model of the Spring Mountains block as proposed by Wernicke (1992). These data have important implications not only for the tectonic evolution of the area, but also for hydrogeologic studies. The LVVSZ appears to affect ground-water chemistry in the valley (Lyles and Hess, 1988) and thus may control ground-water flow. These data allow us to determine the location and geometry of these concealed structures beneath the valley floor.

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