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U.S. GEOLOGICAL SURVEY BULLETIN 1787–J, K

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EVOLUTION OF SEDIMENTARY BASINS—UINTA AND PICEANCE BASINS

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Chapter J

Upper Cretaceous and Paleogene Sedimentary Rocks and Isotopic Ages of Paleogene Tuffs, Uinta Basin, Utah

By BRUCE BRYANT, C. W. NAESER, R. F. MARVIN, and H. H. MEHNERT

A multidisciplinary approach to research studies of sedimentary rocks and their constituents and the evolution of sedimentary basins, both ancient and modern

U.S. GEOLOGICAL SURVEY BULLETIN 1787

EVOLUTION OF SEDIMENTARY BASINS—UINTA AND PICEANCE BASINS

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CONVERSION FACTORS FOR SOME SI METRIC AND U.S. UNITS OF MEASURE

To convert from	To	Multiply by
Feet (ft)	Meters (m)	0.3048
Miles (mi)	Kilometers (km)	1.609
Pounds (lb)	Kilograms (kg)	0.4536
Degrees Fahrenheit (°F)	Degrees Celsius (°C)	Temp °C=(temp °F–32)/1.8

Upper Cretaceous and Paleogene Sedimentary Rocks and Isotopic Ages of Paleogene Tuffs, Uinta Basin, Utah

By Bruce Bryant, C.W. Naeser, R.F. Marvin, *and* H.H Mehnert

Abstract

Reconnaissance geologic mapping and isotopic dating of tuff beds in the Uinta basin of Utah show that Lake Uinta probably persisted into late Eocene time in the area between Duchesne and Strawberry Reservoir. In the area east of Duchesne, contemporaneous deposition occurred on mudflats that were occasionally inundated by lake waters. The sediments deposited in Lake Uinta form the Green River Formation, and those deposited adjacent to Lake Uinta form the Uinta Formation. North of Lake Uinta and west of Duchesne, deposition of quartz sand from the Uinta Mountains produced rocks here included in the Duchesne River Formation.

Uplift of the central Uinta Mountains in late Eocene time produced a flood of quartz sand from the north that was deposited across the axis of the basin to form the Duchesne River Formation in the central and eastern parts of the basin. West of the Duchesne River, the Uinta Mountains continued to be the source of the alluvial-facies sediments that were deposited on top of deposits of Lake Uinta.

In the central Uinta Mountains, relief decreased such that siltstones and tuffs were deposited in the area of the present-day surface structural axis of the basin, quite close to the Uinta Mountains, to form the Lapoint Member of the Duchesne River Formation. Continued uplift of the central and western Uinta Mountains in Oligocene time led to truncation of upturned beds of the lower part of the Duchesne River at the northern margin of the basin and deposition of the Starr Flat Member of the Duchesne River Formation, which consists of boulder to pebble conglomerate and sandstone at the edge of the mountains but grades southward into dominantly sandstone. The Starr Flat Member cannot be distinguished from the overlying Bishop Conglomerate with certainty.

Tuffs interbedded with Paleogene sediments from 24 localities in the Uinta basin were dated by using isotopic methods. Potassium-argon ages of biotite and fission-track ages of zircon indicate that lacustrine deposition in the western part of the basin continued until latest Eocene time, perhaps to 37.6 Ma. Ages determined for tuffs near the type Duchesnean faunal locality in the eastern part of the basin near the base of the Lapoint Member range from 35 to 37 Ma, or very earliest Oligocene. An unconformity at the base of the Starr Flat Member at the northern margin of the basin has been dated as early Oligocene and indicates that a pulse of uplift occurred in the southern part of the Uinta Mountains at that time. Some of the isotopic ages from the Starr Flat Member of the Duchesne River are approximately the same as ages obtained in studies of the Bishop Conglomerate on the southern flank of the eastern Uinta Mountains; mild tectonism probably continued in the area of the Uinta basin into the middle of Oligocene time, a time during which the area of the Uinta Mountains was tectonically quiescent.

Many of the biotite potassium-argon and zircon fission-track ages are discordant. In the tuffs for which the biotite ages are older than the zircon ages, this discordance may result from minor amounts of detrital Precambrian biotite; in tuffs for which the biotite ages are younger, it may result from alteration of the biotite. The fission-track ages generally agree with the stratigraphic order of the samples, but their large analytical uncertainty makes them less than ideal for precise geochronological work.

INTRODUCTION

The Uinta basin is in northeastern Utah, at the northern margin of the Colorado Plateau and south of the east-trending Uinta Mountains (fig. 1). During latest Cretaceous and Paleogene time, as much as 6 km of rocks was deposited in alluvial and lacustrine environments in the deepest part of the basin adjacent to the

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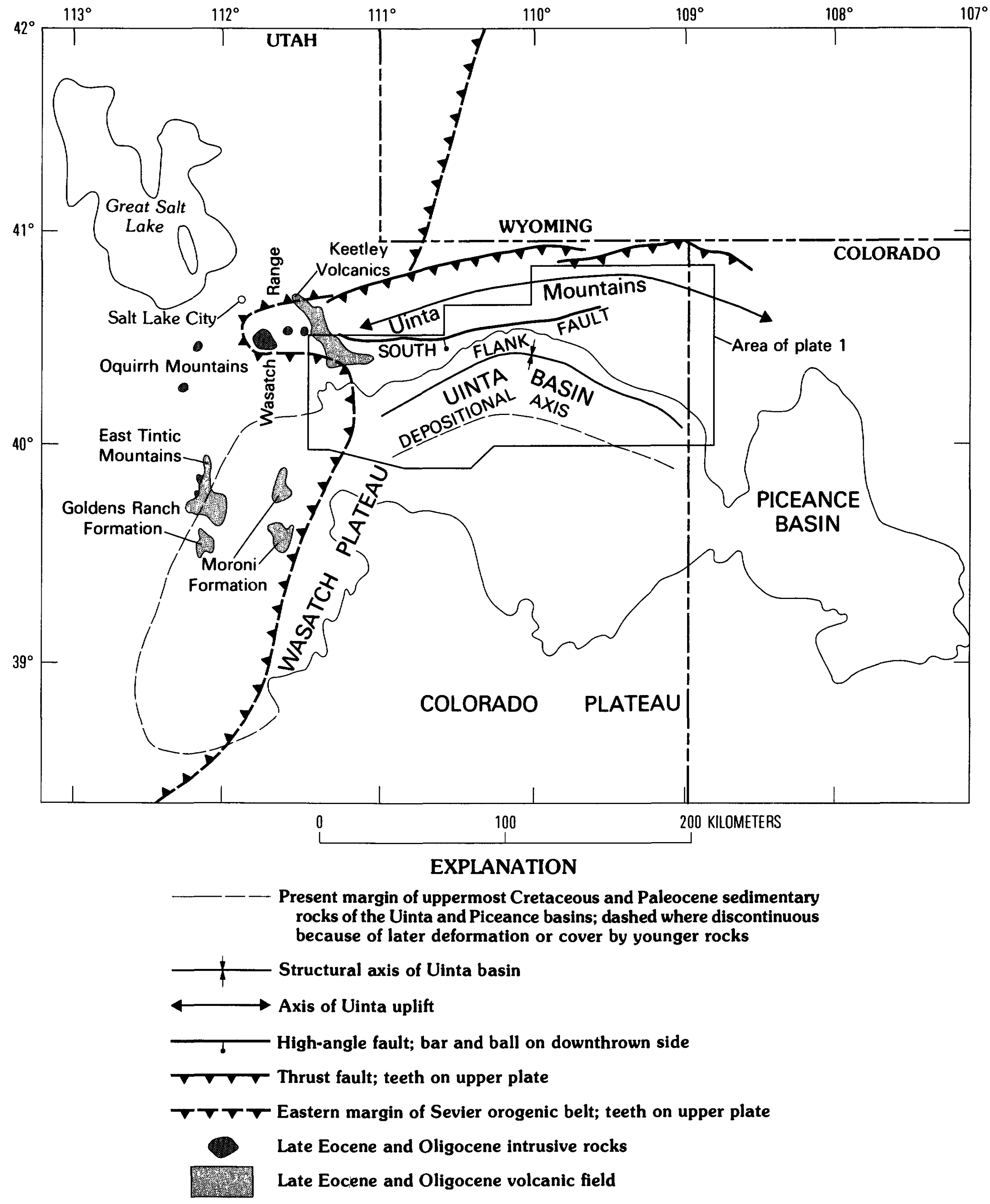

Figure 1. Tectonic setting of Uinta basin, northeastern Utah. Area of plate 1 also shown. Modified from King and Beikman (1974); depositional axis from Johnson (1985) and Ryder and others (1976).

Uinta Mountains. The depositional axis of the basin, along which the most continuous sections of lacustrine rocks occur, is a few kilometers south of the structural axis, along which the thickest deposits are found (Ryder and others, 1976; Johnson, 1985). The axis of a gentle syncline coincides with the location of the thickest deposits and is close to the subsurface position of a thrust fault that was active during part of the depositional history of the basin. The main part of the basin trends east, parallel with the Uinta Mountains, and is about 200 km long and 80 km wide. West of the Uinta Mountains, the basin extended about 150 km southwest and overlapped the margin of the Sevier orogenic belt. Subsequent deformation and erosion has destroyed part of the stratigraphic record in that area, and the deposits preserved are only about 2 km thick.

Reconnaissance geologic mapping of the Salt Lake City 1° × 2° quadrangle by Bryant (in press) resulted in a revised geologic map of the western Uinta basin (plate 1). This map, in conjunction with fission-track dating of zircon by Naeser and potassium-argon (K-Ar) dating of biotite by Marvin and Mehnert from tuff beds in the Paleogene sedimentary sequence in the basin, furnishes a new geochronologic framework with which to interpret stratigraphic relations in the basin.

A part of the northern margin of the Uinta basin and segments of its depositional and structural axes are within the area of the Salt Lake City quadrangle. In the center of the basin, alluvial-facies rocks of the Duschesne River and Uinta Formations overlie lacustrine-facies rocks of the Green River Formation. Along the northern flank of the basin, one or the other of these formations intertongues with the Green River Formation in the subsurface. In the western part of the basin, west of Strawberry Reservoir, the Duschesne River Formation intertongues with the Green River Formation and the marginal part of the lacustrine facies of the Green River is exposed.

Acknowledgments.—Reconnaissance geologic mapping in the eastern part of the basin by Rowley and others (1985) in the Vernal 1° × 2° quadrangle and by Weiss and others (in prep.) and Witkind and Weiss (1985) in parts of the Price 1° × 2° quadrangle provides data that allowed extension of our geochronology study to the east and south of the Salt Lake City quadrangle. T.D. Fouch and W.B. Cashion shared their knowledge of the stratigraphy of the Uinta basin through office and field conferences. D.M. Cheney made the zircon and biotite separations.

PREVIOUS WORK

Stratigraphic studies by Bradley (1931), Peterson and Kay (1931), and Kay (1934) provided the first detailed information and interpretation of rocks in the north-central Uinta basin. The lacustrine rock sequence includes a basal lacustrine phase, tongue of Wasatch Formation, second lacustrine phase, delta facies, oil shale facies, and the barren and saline facies of the Green River Formation (Bradley, 1931). The overlying alluvial-facies sequence originally was designated the Uinta Formation (Marsh, 1871) and later subdivided into the Uinta and Duchesne River Formations (Peterson and Kay, 1931; Kay, 1934). The vertebrate fauna of the alluvial-facies rocks was described by Marsh (1871), Osborn (1855), Douglass (1914), Peterson (1931, 1932, 1934), and Burke (1934).

Exploration for and development of oil and gas fields in the north-central Uinta basin since 1950 have made much subsurface data available. These subsurface data, as well as later studies, have resulted in a number of modifications to the original nomenclature within the Green River Formation. Picard (1955, 1959) substituted the green-shale facies for the delta facies of Bradley (1931) and the black-shale facies for Bradley's basal lacustrine phase. Dane (1954, 1955) put the saline facies and overlying lacustrine rocks into the Uinta Formation because of their demonstrable time equivalence to rocks of the Uinta Formation in the eastern part of the Uinta basin based on surface stratigraphic studies. He extended the terms Evacuation Creek and Parachute Creek Members of the Green River Formation from the eastern part of the Uinta basin to the western part. Subsequently, the term Evacuation Creek Member was abandoned (Cashion and Donnell, 1974) after studies showed that the upper part of the Parachute Creek Member at its type locality to the east in the Piceance basin (fig. 1) is the same lithologic and stratigraphic unit as the Evacuation Creek Member at its type locality in the Uinta basin (Cashion and Donnell, 1974). Ryder and others (1976) integrated surface and subsurface studies to analyze the early Tertiary sediments in the basin in terms of open-lacustrine, marginal-lacustrine, and alluvial facies, and they extended the term Green River Formation upsection to include the sandstone and limestone facies near Duschesne, which previously had been included in the Uinta Formation by both Bradley (1931) and Dane (1954). Because a continuous section of lacustrine rock extends down to the base of the Flagstaff Limestone in the center of the basin, the Flagstaff is now considered a member of the Green River Formation (Fouch, 1976; Ryder and others, 1976), and thus Paleocene as well as Eocene beds are included in the Green River Formation.

Much less work has been done on the alluvial-facies rocks comprising the Duchesne River and Uinta Formations. Kay (1934) divided the Duchesne River Formation into three "horizons." In a sedimentological study of the Duchesne River Formation east of the Duchesne River, Warner (1966) divided the formation into a major bentonite member overlying a minor bentonite member, both of which occur in an eastern mudstone facies and a western sandstone facies. Andersen and Picard (1972) proposed four formal members of the Duschesne River Formation (in ascending order), the Brennan Basin, Dry Gulch Creek, Lapoint, and Starr

Flat. Although they stated that these members are traceable throughout the outcrop area of the Duchesne River Formation, they published no map showing their distribution.

The Bishop Conglomerate, named by Powell (1876) from a type locality north of the Uinta Mountains in Wyoming, was interpreted by Bradley (1936) as a pediment gravel deposited on a widespread erosion surface on the north flank of the Uinta Mountains. It has been mapped over large areas in the eastern Uinta Mountains (Rowley and others, 1985), and more recent studies support Bradley's interpretation (Hansen, 1984). The Bishop Conglomerate has been tentatively mapped as overlying the Uinta and Duchesne River Formations and capping high hills such as Tabby Mountain and Dry Mountain at the northern margin of the western part of the basin (Huddle and McCann, 1947; Huddle and others, 1951) and on the high ridges surrounding the head of Currant Creek in the northeastern part of the basin (Bissell, 1952).

STRATIGRAPHY

The most complete stratigraphic sections of Upper Cretaceous and Tertiary rocks in the Uinta basin are deeply buried near the depositional axis, where deposition of lacustrine rocks persisted for the longest period of time. This part of the basin has been penetrated to depths as great as 6 km by wells drilled for oil and gas, and some of the stratigraphic relations revealed in these data have been summarized by Ryder and others (1976) and Fouch (1981). Much of the surface of the Uinta basin is underlain by younger alluvial-facies rocks. Along the northern margin of the basin, the alluvial-facies rocks represent a period of time from the inception of the basin in late Campanian-early Maestrichtian time to the end of basin deposition in middle Oligocene time. Angular unconformities in basin-margin rocks indicate that some geologic time is not represented and that basin formation was concommitant with uplift of the Uinta Mountains to the north (Walton, 1944).

Currant Creek Formation

The oldest stratigraphic units discussed in this report are the North Horn and Currant Creek Formations. The Currant Creek Formation (Walton, 1944) crops out only along the north side of the Uinta basin, from near Strawberry Reservoir on the west nearly to Dry Mountain on the east. In its western part, it is composed of thick conglomerate rich in pebbles, cobbles, and boulders of rocks resembling those of the Oquirrh Group exposed in the Charleston thrust sheet just west of the map area and clasts of Late Proterozoic quartzite derived from the northwest (Isby and Picard, 1983). To the east, the Currant Creek contains more sandstone and siltstone and finer grained conglomerate and is overlapped by younger rocks. Although the Currant Creek appears to be concordant with the underlying Mesaverde Formation of Late Cretaceous age, the Mesaverde is partly cut out in an easterly direction such that its upper coal-bearing part does not extend to the Duchesne River (Lupton, 1912).

Paleocurrent studies indicate that the Currant Creek was deposited by streams flowing south to southeast, and a marked decrease in grain size from west to east suggests that the source of the sediments was northwest of the outcrop belt of the Currant Creek (Isby and Picard, 1983).

The base of the Currant Creek has yielded palynomorphs and ostracods indicative of late Campanian or early Maestrichtian age. At Red Creek, the underlying Mesaverde contains palynomorphs of Coniacian or Santonian age, and the unconformity at the base of the Currant Creek represents a hiatus of 9–15 m.y. (Nichols and Bryant, 1986). Only the lowermost 100 m of the Currant Creek has been dated, and deposition of the Currant Creek may have extended into Paleocene time.

The dip of beds within the Currant Creek decreases up section, indicating that the margin of the basin was being deformed during deposition. This deformation probably was related to uplift of the Uinta Mountains.

North Horn Formation

The North Horn Formation consists of conglomerate, sandstone, siltstone, mudstone, and a few beds of limestone; it crops out in the Diamond Fork region in the western part of the Uinta basin and along the southern margin of the basin where it forms the basal deposit in the western Uinta basin (Fouch and others, 1983). Ryder and others (1976) extended the term North Horn to include the dominantly alluvial-facies rocks underlying the lacustrine-facies rocks of the interior of the basin. The North Horn contains local beds of claystone, sandstone, and argillaceous limestone that accumulated in paludal and marginal-lacustrine environments (Ryder and others, 1976, p. 510). On Soldier Creek, west of Soldier Summit, and on Diamond Fork, the North Horn contains thick beds of cobble conglomerate that represent a proximal alluvial facies. The basal part of the North Horn is late Campanian(?) to early Maestrichtian in age in Price Canyon south of the area of plate 1 and has yielded dinosaur remains of latest Cretaceous age on the Wasatch Plateau (Spieker, 1946, p. 134–135). Near Soldier Summit, the basal part contains Cretaceous palynomorphs of an unspecified zone (Newman, 1974). No paleontological information has been obtained from the North Horn farther to the northwest in the western Uinta basin, but the wedging out of the North Horn beneath the Flagstaff Member of the Green River Formation in the Wasatch Mountains (Baker, 1976) suggests that the basal beds of the North Horn near the margin of the basin are much younger and are Paleocene in age.

The North Horn Formation is correlative with the Currant Creek Formation in the northwest part of the basin, but its top is defined by lacustrine rocks of the Flagstaff Member of the Green River Formation rather than by alluvial rocks of the Duchesne River Formation. The term North Horn is not applicable at the northern margin of the basin because the stratigraphic section is composed entirely of alluvial-facies rocks.

Green River Formation

The Green River Formation consists of lacustrine limestone, dolomite, marlstone, shale, sandstone, oil shale, and tuff. The rock types and mineralogy of the Green River have been described in a number of published reports, including those by Bradley (1931), Picard (1955), and Ryder and others (1976). Along the southern and western margins of the basin, the North Horn Formation is overlain by lacustrine limestone of the Flagstaff Member at the base of the Green River Formation. Along the western margin of the basin, the Flagstaff Member contains interbeds of gray to moderate-red shale, siltstone, and conglomerate that are, in part, tongues of alluvial-facies rocks (Baker, 1976). At the western margin of the basin, the Flagstaff Member is overlain by shale and siltstone of the Green River Formation. In the interior of the basin, however, data from wells show that the Flagstaff merges with and becomes indistinguishable from the overlying thick mass of lacustrine-facies rocks of the Green River Formation (Fouch, 1976; Ryder and others, 1976). In many areas, both in the subsurface and on the surface, alluvial-facies red sandstone, siltstone, and mudstone of the Eocene Colton Formation separate the Flagstaff Member of the Green River Formation from the overlying lacustrine rocks of the main body of the Green River Formation (Fouch, 1976; Ryder and others, 1976), but this tongue of alluvial-facies rocks of the Colton fingers out in surface exposures in the western Uinta basin just south of the Salt Lake City quadrangle. No ages have been obtained for the Flagstaff in the map area, but elsewhere paleontologists have dated it as late Paleocene and early Eocene in age (La Roque, 1960; Newman, 1974; Ryder and others, 1976, p. 497).

The main body of the Green River Formation in the western Uinta basin has been subdivided in various ways (Bradley, 1931; Dane, 1954; Picard, 1959; Moussa, 1969), but we follow the suggestions of Ryder and others (1976) and include in the Green River Formation rocks deposited predominantly in a lacustrine environment and avoid formal stratigraphic names for subdivisions above the Flagstaff Member. Study of surface sections and information from drill holes shows that the rock types in the basin form a complex pattern, and application of stratigraphic names to lithologic packages may be misleading. In the subsurface, however, three marker beds in the Green River Formation help correlate various sections: the carbonate marker unit, the middle marker, and the Mahogany oil shale bed (Ryder and others, 1976, p. 497). The lower two beds crop out on the surface south of the Salt Lake City quadrangle (Ryder and others, 1976, p. 497), and the Mahogany bed has been traced in the surface and subsurface throughout the Piceance and Uinta basins (Cashion, 1967; Cashion and Donnell, 1972; Fouch and Cashion, 1979; Fouch, 1981) and crops out in the southern part of the map area (plate 1).

In the Salt Lake City quadrangle, an oil shale unit in the upper part of the Green River Formation, about two-thirds of the distance above the base, is as thick as 5 m and contains a tuff bed 3 cm thick. This bed was not traced west of sec. 11, T. 10 S., R. 6 E., but probably thins and becomes less distinct toward the northwest margin of the basin. The oil shale unit is identified as the Mahogany ledge because of its regional continuity and stratigraphic position in relation to two distinctive, widespread tuff beds (W.B. Cashion, oral commun., 1981). The Mahogany oil shale bed lies within the Mahogany ledge and is a stratigraphic time marker throughout most of the basin. The upper part of the Green River Formation containing the Mahogany ledge has been called the Parachute Creek Member in this part of the basin (Dane, 1954, 1955), but the term is not used in this report. In the west-central Uinta basin, strata above and below the Mahogany ledge are lithologically different than the type Parachute Creek Member (W.B. Cashion, written commun., 1985).

In the west-central part of the basin, rocks in the upper part of the continuous lacustrine sequence were assigned to a saline facies of the Green River Formation by Bradley (1931). They include limestone, marl, and shale and are characterized by abundant molds of saline mineral crystals that give them a distinctive mottled texture. Some 24 authigenic sodium minerals have been identified in core from a well north of Duchesne (Dyni and others, 1985). Dane (1954, 1955) included these rocks in the Uinta Formation because of their time equivalence with rocks of the Uinta Formation farther to the east. Picard (1959) included most, but not all, of the saline-facies rocks in the Green River Formation. Following Ryder and others (1976) and Dyni and others (1985), we assign the upper lacustrine rocks to the Green River Formation.

In the lower part of the sequence herein mapped as saline facies, a zone containing numerous beds of tuff was called the upper tuff zone of the Evacuation Creek Member of the Green River Formation by Dane (1954, 1955). In Indian Canyon, about 20 m above the site of the Gulf Oil Company Indian Canyon No. 1 well in the north part of sec. 12, T. 6 S., R. 7 W., Dane measured about 65 m of section in the tuff zone, the base of which is marked by a 1-m-thick bed of biotitic tuff. Bradley (1931) and Dyni and others (1985) believed that this tuff is the base of the saline facies, whereas Dane placed the base of the saline facies 65 m higher in the section at the top of the tuff zone. We place the base of the saline facies about 15 m above that tuff, but because the base is gradational with underlying rocks, differences in placement of the base of the saline facies by different workers are to be expected.

The tuff zone is well exposed along Lake Canyon and along the right and left forks of Indian Canyon. The valley bottom of Lake Canyon must be almost parallel with bedding because the tuff zone crops out on the lower part of the valley sides from its mouth to the south edge of the Salt Lake City quadrangle, a distance of about 20 km. The tuff zone crops out along the lower valley side of the Strawberry River in secs. 7–8 and 11–15, T. 4 S., R. 6 W.

At the top of the tuff zone as delineated by Dane (1955), a tuff about 1 m thick overlies a ledge-forming, kerogen-rich limestone that shows the mottled texture characteristic of the saline-facies rocks. This tuff was mapped photogrammetrically as horizon 9 by Ray and others (1956) and has been recognized as far west as south of Strawberry Reservoir, where Moussa (1969) mapped it as the base of his saline facies of the Uinta Formation. Numerous tuff beds occur in the 60 m of section above this marker.

About 30 m above the tuffs is the base of a zone of yellowish-gray-weathering rock called mealstone that is composed of oil shale and contains abundant molds and pseudomorphs of saline minerals (Dyni and others, 1985). In many valleys, the yellowish-gray-weathering rock can be traced for miles along the canyon walls.

The upper part of the saline facies contains fewer beds showing mottled textures and more beds of chert and sandstone than does the lower part, and it grades upward into the sandstone and limestone facies of the Green River Formation.

East of Indian Canyon, saline-facies rocks intertongue with sandstone- and limestone-facies rocks that in turn intertongue with alluvial-facies rocks of the lower part of the Uinta Formation (Dane, 1954, 1955). The saline-facies rocks extend westward in approximately the same stratigraphic interval, as judged by their relation to the tuff zone, to as far west as the Wasatch County-Utah County line 10 km southwest of Strawberry Reservoir where they interfinger with alluvial-facies rocks of the Duchesne River Formation. In Tps. 4, 5 S., Rs. 9, 10 W., the saline facies is much thinner than elsewhere in the map area and is represented by a section rich in sandstone.

The saline facies is overlain by the sandstone and limestone facies, which was named and assigned to the Uinta Formation by Dane (1954, 1955). Its lower contact is at the upper limit of the mottled texture characteristic of the saline facies and at the first occurrence of fish remains. Rocks of the sandstone and limestone facies grade upward into rocks of the alluvial facies. In the Salt Lake City quadrangle, west of Duchesne and at Red Creek, the upper contact of the saline facies is where Dane described the contact between the lacustrine and alluvial facies. East of Duchesne, the lacustrine rocks of the sandstone and limestone facies intertongue with rocks of the alluvial facies of the lower part of the Uinta Formation and with rocks of the saline facies of the Green River Formation. Individual beds of white lacustrine limestone can be traced for 20–30 km to the east (Ray and others, 1956) into the predominantly alluvial section of the Uinta Formation, which is characterized by greenish-gray to light-gray and grayish-red siltstone and claystone. The lacustrine-facies rocks of the Green River Formation are gradational with the alluvial-facies rocks of the Uinta Formation.

To the west, both along and south of Strawberry River between Avintaquin Canyon and Willow Creek, rocks of the sandstone and limestone facies interfinger with a unit that is predominantly sandstone. West of Willow Creek, this sandstone interfingers with rocks of the sandstone and limestone facies, which contains more limestone than the same unit near Duchesne, as well as a few thin beds of oil shale. South of Strawberry Reservoir, rocks of the sandstone and limestone facies interfinger with rocks of the alluvial facies to the north, along the former margin of Lake Uinta.

Along Strawberry River, Beaver Canyon, Slab Canyon, and lower Avintaquin and Timber Canyons, most rocks are a light-grayish-yellow and yellowish-brown, thick- to thin-bedded calcareous sandstone that contains beds of sandy limestone, shale, and tuff. This calcareous sandstone grades southward into shale and limestone of the Green River Formation below the saline facies. A tongue of sandstone extends down the Strawberry River to Sams Canyon, where it enters the subsurface. The western extent of the sandstone body is not well defined because it is poorly exposed and gradational with rocks of the sandstone and limestone facies to the west. A tongue of sandstone extends into the upper part of the sandstone and limestone facies unit in R. 11 W., south of Strawberry Reservoir. In the parts of the sandstone unit stratigraphically equivalent to the saline facies, mottled textures characteristic of saline facies occur in limestone interbeds. The generally even and parallel bedding in the sandstone suggests that it was waterlain, and the sandstone probably represents a delta deposit of quartz sand laid down by a major drainage system from the north or northwest.

Uinta Formation

The Uinta Formation overlies the Green River Formation in the eastern and east-central parts of the Uinta basin. Most previous studies of the Uinta Formation have been by vertebrate paleontologists because the formation contains the distinctive Uintan fauna of late Eocene age.

In the Vernal 1°×2° quadrangle, in the eastern part of the Uinta basin, the Uinta Formation has been divided into a lower unit characterized by sandstone and minor shale in its lower part and by minor sandstone and channel sandstone beds scattered in a dominantly claystone and sandstone section in its upper part, and an upper unit of variegated shale, mudstone, claystone, and minor sandstone (Rowley and others, 1985). The upper unit and part of the lower unit can be traced westward into the Salt Lake City quadrangle almost to Duchesne, at which point the lower unit grades westward into the sandstone and limestone facies of the Green River Formation. The widespread occurrence, from near Duchesne almost to the Duchesne-Uintah County line, of thin beds of lacustrine limestone in the variegated siltstone and claystone of the lower unit (Ray and others, 1956) indicates that the rocks of the Uinta Formation in the area were deposited on mudflats that were inundated, from time to time, by the waters of Lake Uinta. The lower part of the lower unit does not occur in the Salt Lake City quadrangle because it grades laterally into stratigraphically equivalent rocks of the sandstone and limestone facies and saline facies of the Green River Formation. The upper unit of the Uinta Formation in the east-central part of the map area differs from that in the Vernal quadrangle to the east in that it is rich in gray clay and has a few beds of paludal or lacustrine limestone.

Near the top of the upper unit of the Uinta Formation in sec. 33, T. 2 S., R. 4 W., a limestone bed containing ostracods and fish remains is interbedded with papery shale, gypsiferous limestone, other thin limestone beds, and sandstone. In sec. 30, T. 2 S., R. 2 W., a limestone containing ostracod and fish remains crops out about 10 m below the top of the Uinta Formation. Rocks equivalent to this stratigraphic interval can be traced westward to Blacktail Mountain, where they consist of greenish-gray siltstone and clay that contrast with the pale-red sandstone and moderate-red siltstone above and below. Along the Duchesne River north of Duchesne, gray and greenish-gray clay typical of the Uinta Formation interfingers to the west with gray, yellowish-gray, and grayish-red quartz sandstone similar to that in the overlying Duchesne River Formation. A few kilometers farther to the west, these sandstone beds become dominant and the section is included in the Duchesne River Formation (plate 1).

Paleocurrent and petrographic data indicate the Uinta Formation was deposited by streams flowing into the basin from the east and southeast (Stagner, 1941; Bruhn and others, 1986).

Duchesne River Formation

In the central and eastern Uinta basin, red sandstone, shale, and siltstone in the upper part of the section originally called Uinta Formation were later given the name Duchesne River Formation (Peterson and Kay, 1931; Kay, 1934). These rocks lap onto the depositional axis of the basin (fig. 1, plate 1). The Duchesne River Formation originally was divided into the Randlett, Halfway, and Lapoint "horizons," but these units were not well defined or mapped. Warner (1966) subdivided the Duchesne River into two units, a lower minor bentonite member and an upper major bentonite member. Both members occurred in an eastern muddy facies and a western sandy facies. Andersen and Picard (1972) divided the Duchesne River Formation into four members, and, although they described them thoroughly and gave detailed stratigraphic sections, they did not show their map distribution. Bryant (this paper) and Rowley and others (1985) mapped in reconnaissance the outcrop area of the Duchesne River and were able to map the four members of Andersen and Picard (1972) in the central and eastern parts of the basin, but Bryant was not able to extend these units into that part of the basin west of Rock Creek (plate 1).

The basal beds of the Duchesne River Formation intertongue with the Uinta Formation in the eastern and central Uinta basin (Andersen and Picard, 1972; Rowley and others, 1985). Nevertheless, the contact between Duchesne River and Uinta rocks is quite well defined in the map area east of the Duchesne River. Along that contact, yellowish-brown to moderate-red sandstone of the Brennan Basin Member of the Duchesne River Formation rests on a gray claystone sequence of the Uinta Formation that contains some thin-bedded, yellowish-brown sandstone beds.

Along the Duchesne River, between the town of Duchesne and Rock Creek, thick beds of quartz sandstone characteristic of the Duchesne River Formation intertongue eastward with gray claystone of the Uinta Formation over a distance of about 6 km.

Farther to the west, this entire interval resembles the Duchesne River and was mapped as undivided Duchesne River.

Paleocurrent studies show that sediments of the Duchesne River were deposited by streams flowing southward from the Uinta Mountains (Warner, 1966; Andersen and Picard, 1972, 1974; D.W. Andersen, written commun., 1980).

Brennan Basin Member

The Brennan Basin Member of the Duchesne River Formation overlies the Uinta Formation in the east-central and eastern parts of the basin (plate 1). It consists of moderate-red, grayish-red, reddish-brown, yellowish-brown, and yellowish-orange sandstone and lesser amounts of reddish-brown siltstone. Although it contains a few beds of pebble conglomerate on Asphalt Ridge at the northeastern margin of the basin, the lower part is predominantly light gray, yellowish gray, and pink conglomerate (Rowley and others, 1985). About 3 km northeast of Randlett, Utah, a Uintan vertebrate fauna was collected near the base of the unit. (See summary in Andersen and Picard, 1972). Near Asphalt Ridge, the Uinta, Green River, and underlying Wasatch Formations pinch out, and the Brennan Basin Member overlies Mesozoic rocks and contains more conglomerate than to the southwest (Rowley and others, 1985). The Brennan Basin Member is tilted almost as steeply as the underlying Mesozoic rocks, indicating that most deformation along the basin margin occurred after deposition of the Brennan Basin Member. West of Rock Creek, the Brennan Basin Member loses its identity, and rocks of this part of the section are mapped as part of the undivided Duchesne River Formation. Along the east side of Rock Creek Valley, the base of the member cuts upsection in a northerly direction such that the contact is about 200 m higher in the section 10 km up Rock Creek than it is at its mouth.

Dry Gulch Creek Member

The Dry Gulch Creek Member consists of moderate-red and grayish-red sandstone and siltstone and gray claystone and contains a greater proportion of fine-grained rocks than the Brennan Basin Member. In the eastern part of the basin, it contains some conglomerate where it is exposed close to the basin margin (Rowley and others, 1985). West of Roosevelt, sandstones of the Dry Gulch Creek are yellower and siltstones grayer than to the east. Some of the gray claystone beds contain euhedral biotite grains and are interpreted as altered tuffs; many of the other claystone beds probably represent altered tuffaceous deposits.

The Dry Gulch Creek Member overlies and interfingers with the Brennan Basin Member and has been mapped over a distance of 80 km from Asphalt Ridge to Rock Creek (plate 1). West of Roosevelt, the basal beds interfinger downsection to the west. On the east side of Rock Creek, the base cuts upsection in a northerly direction.

Lapoint Member

In the east-central part of the basin, the Lapoint Member consists predominantly of gray siltstone and claystone and contains some beds of orangish-gray sandstone and a few beds of limestone. Light- to medium-gray or blue-gray bentonite beds, some of which contain euhedral biotite crystals, are numerous. These beds are altered tuffs, and much of the claystone may be altered tuff or tuffaceous clay and silt. In the eastern part of the basin and near the Duchesne-Uintah County line on the flank of the Uinta Mountains, many of the siltstone and claystone beds are reddish brown rather than gray.

The Lapoint Member overlies the Dry Gulch Creek Member and crops out along the structural axis of the Uinta basin between Asphalt Ridge and Rock Creek. East of Bluebell, basal beds of the Lapoint interfinger northward with the Dry Gulch Creek Member. West of Mountain Home, the Lapoint Member interfingers westward with the Dry Gulch Creek Member. At the basin margin northwest of Vernal, the Lapoint Member interfingers northward with the overlying Starr Flat Member. Near the Uintah-Duchesne County line on the north margin of the basin, the Lapoint Member interfingers westward into sandstones of the Starr Flat Member.

Starr Flat Member

The Starr Flat Member ranges from pale-red, moderate-red, reddish-brown, and grayish-red boulder conglomerate, sandstone, and minor claystone near the mountains to yellowish-gray sandstone and minor gray and greenish-gray sandy claystone near the axis of the Uinta basin. Rowley and others (1985) included much of the proximal-facies conglomerate of the Starr Flat Member in the Bishop Conglomerate at the northern margin of the Uinta basin. Proximal-facies rocks of the Starr Flat Member are not exposed in most places; however, on Little Mountain in the eastern part of the basin, sandstone, siltstone, and minor conglomerate mapped as Bishop (Rowley and others, 1985) are overlain conformably by conglomerate containing a few sandstone and siltstone interbeds also mapped as Bishop by the authors. We call all these rocks Starr Flat Member, and if the Bishop Conglomerate is in this area, we believe it caps the flat-topped ridge crests and has an indeterminant contact with the Starr Flat Member.

The Starr Flat Member overlies the Lapoint Member along the northern margin of the basin and overlaps Mesozoic, Paleozoic, and Precambrian rocks at the margin of the Uinta uplift from the Duchesne-Uintah County line to Rock Creek (plate 1). West of Rock Creek, conglomerate that forms the proximal facies of the Starr Flat Member underlies Tabby Mountain, Dry Mountain and Raspberry Knoll and overlies finer grained, more steeply dipping rocks mapped as undivided Duchesne River Formation. Within a few kilometers of the basin margin, the conglomerate grades to sandstone that dips concordantly with the underlying part of the Duchesne River. Rowley and others (1985) mapped the Bishop Conglomerate as overlying the Starr Flat Member, and although we show the Bishop Conglomerate as overlying the Starr Flat Member on the high, flat-topped hills on either side of the Uinta River, we are uncertain if there is any significant difference in lithology and age between the Starr Flat Member and the Bishop Conglomerate. The flat-topped hills apparently represent constructional surfaces at the top of an alluvial-fan complex deposited by streams emanating from the Uinta Mountains. West of the Yellowstone River, the remnants of these surfaces are more eroded, and we do not map any Bishop; conglomerate in that area has been called Bishop by previous workers (Huddle and McCann, 1947; Huddle and others, 1951), who pointed out the uncertainty of the designation.

Near Little Mountain in the eastern part of the basin, the Starr Flat Member forms an isolated alluvial-fan complex that unconformably overlies moderately dipping rocks of the Brennan Basin Member (Rowley and others, 1985).

Undivided Duchesne River Formation

West of Rock Creek and the Duchesne River, rocks of the Duchesne River Formation were not subdivided; north of Duchesne, part of the Duchesne River interfingers with the upper part of the Uinta Formation. West of Strawberry Reservoir, the Duchesne River interfingers with the Green River Formation, and at the margin of the basin the entire Green River Formation grades into alluvial-facies rocks of the Duchesne River Formation, and the alluvial-facies rocks that lie north of the pinchout of the Flagstaff Member of the Green River are mapped as an unnamed unit (map unit Ta) that may contain beds of Paleocene, Eocene, and Oligocene age. To the south, in the Diamond Fork area, the rocks overlying the Flagstaff Member of the Green River Formation and underlying a unit of sandstone, conglomerate and tuff mapped as the Moroni Formation (Witkind and Weiss, 1985) are included in the undivided Duchesne River Formation rather than in the Green River and Uinta Formations. Rocks in that area were mapped by Baker (1976) as Green River Formation and are predominantly interbedded red and greenish-gray conglomerate, siltstone, and claystone, and minor limestone that we interpret to be dominantly alluvial facies but contain a few tongues of lacustrine rock. These rocks do not differ much from those mapped by Baker as Uinta Formation. Young (1976) and Witkind and Page (1983) used the term undifferentiated Green River and Colton Formation to identify the rocks Baker mapped as Green River in the Diamond Fork area. Because the Colton Formation is defined as a tongue of alluvial-facies rock between the Flagstaff Member of the Green River and the main body of the Green River (Fouch, 1976), the application of the term Colton to alluvial-facies rocks overlying the Flagstaff Member but not overlain by the main body of the Green River is inappropriate.

The undivided Duchesne River Formation contains abundant conglomerate near the basin margin from Coop Creek to the Duchesne River. The gradation from predominantly sandstone and siltstone to conglomerate and sandstone is well exposed in the valley of Currant Creek, north of U.S. Highway 40. This gradual change in the almost flat lying rocks occurs over a distance of 15 km along that valley northward toward the basin margin. To the east, on the lower flanks of Tabby and Dry Mountains, undivided Duchesne River dips south as steeply as 30° and is unconformably overlain by conglomerate of the Starr Flat Member that dips 5° toward the center of the basin; this difference in dip indicates that uplift along the margin of the basin occurred after deposition of the undivided Duchesne River and before deposition of the Starr Flat Member. On either side of Dry Mountain, the undivided Duchesne River contains relatively numerous beds of gray clay, some of which contain euhedral biotite crystals. These beds are certainly altered tuffs, and many other beds of gray clay lacking the biotite probably are tuffaceous. East of the Rock Creek drainage, the more steeply dipping, tuffaceous part of the undivided Duchesne River is not exposed. At the margin of the basin in the Asphalt Ridge-Little Mountain area, similar dips in the Brennan Basin Member of the Duchesne River indicate that deformation in that area occurred after deposition of the Brennan Basin Member.

In the Strawberry graben, the undivided Duchesne River Formation contains some biotitic tuff and beds of tuffaceous limestone and conglomerate that contain clasts of volcanic rock. Exposures of these rocks are few, and we do not know if there is a separate unit containing volcanic debris or if the beds containing volcanic detritus are interbedded with nonvolcanic conglomerate. At present, we believe the latter to be the case. In the Red Creek area near Fruitland, the undivided Duchesne River is rich in reddish-brown siltstone and claystone.

The Bishop Conglomerate may be equivalent to part of the Starr Flat Member of the Duchesne River Formation. Some areas previously been mapped as

Bishop in the western part of the basin, such as Tabby Mountain and Dry Mountain, are composed of conglomerate that grades basinward into sandstone. In the central part of the basin, we mapped Bishop as a cap on flat-topped, southward-protruding shoulders of the Uinta Mountains. The conglomerate on those shoulders seldom forms outcrops because it is weakly cemented and commonly is covered by a well-developed soil. On the shoulder of the Uinta Mountains, just west of the Whiterock River, a tuff several meters thick is immediately beneath the rocks shown as Bishop on plate 1.

FAUNAL AGES OF THE DEPOSITS

Dating of Tertiary rocks of the Uinta basin has been based historically on the vertebrate faunas from beds of the alluvial facies. Laterally equivalent lacustrine beds were dated by stratigraphic relations with alluvial beds that contain vertebrate fossils. The main body of the Green River Formation in the Uinta basin was considered to be in the upper part of the lower Eocene and the middle Eocene, the Uinta Formation late Eocene, and the Duchesne River Formation late Eocene and perhaps Oligocene. (For a brief summary of these ages, see Gazin, 1959.)

The age of the Duchesne River Formation has been a subject of controversy. Except for vertebrates collected near the base of the Brennan Basin Member, the fauna of the Duchesne River Formation differs from that of the underlying Uinta Formation of late Eocene age and also from that of the Chadron fauna of basal Oligocene rocks of the High Plains east of the Rocky Mountains. The fauna of the Duchesne River Formation, mostly found in the Lapoint Member in one area east of Lapoint, forms the basis of the Duchesnean faunal zone. Emry (1981) described two new taxa from the Lapoint Member, reviewed previously reported fauna, and concluded that early Chadronian faunas are more similar to those of the Duchesne River Formation than previously thought. He believed that the Duchesnean fauna is correlative with the earliest part of the Chadronian fauna and is early Oligocene in age.

ISOTOPIC AGES OF THE TUFFS

Previous work

Isotopic dating of minerals from tuffs interbedded with the Tertiary rocks of the Uinta basin helps assign ages to the rocks and the mammalian faunas contained in them. Evernden and others (1964), in a worldwide study that included no samples of rocks of the Uinta basin, obtained a K-Ar age of 38.5 Ma from a Duchesnean faunal locality and an age range of 32.4–37.8 Ma from Chadronian faunal localities. (If necessary, previously reported K-Ar ages were corrected for currently accepted decay constants by using table 2 of Dalrymple, 1979.) McDowell and others (1973) reported a biotite K-Ar age of 37.2 Ma for a tuff in the lower member of the Chadron Formation of Oligocene age in South Dakota and a biotite K-Ar age of 40.3 Ma for a tuff at the base of the Lapoint Member of the Duchesne River east of Lapoint from which Duchesnean fauna had been collected.

K-Ar ages for biotite from several tuffs in the Uinta basin were reported by Mauger (1977): 46.0 and 45.6 Ma for the Wavy tuff (loc. 23), 50 m above the Mahogany oil shale in the Mahogany zone; 42.9 Ma for a tuff below the base of the saline facies in Indian Canyon (loc. 22); 42.7 Ma for a bed near the top of the saline facies (all of the above from the Green River Formation); and 38.0 and 35.7 Ma for tuff at the top of the Dry Gulch Creek Member of the Duchesne River Formation (1oc. 15). (All numbered sample localities are shown on plate 1 and described in appendix.) Ages determined by $^{40}Ar/^{39}Ar$ age spectrum techniques (O'Neill and others, 1980) are 46.7 Ma for the Wavy tuff and 47.2 Ma for the Curly tuff, which is about 25 m below the Mahogany oil shale bed in the Green River Formation. Stratigraphic thicknesses between the tuffs and the oil shale bed are from a section measured by W.B. Cashion and J.R. Dyni in Avintaquin Canyon (W.B. Cashion, written commun., 1982).

The exact age of the Oligocene-Eocene boundary has been a subject of controversy. In marine sections, the boundary has been related to the magnetic polarity time scale, and calibration of that time scale is uncertain. Based on a magnetostratigraphic study of a section of early Oligocene rocks in Wyoming, Prothero (1985) concluded that the Eocene-Oligocene boundary must lie between 36 and 37 Ma. This time range has been accepted by the Committee on Geochronology as the standard for the Decade of North American Geology (Berggren and others, 1985). The committee report indicates, however, that Duchesnean fauna existed from 38 to 42 Ma and therefore is entirely Eocene in age and Chadronian fauna existed as long ago as 39 Ma and thus also extended into Eocene time. We suggest that this rather confusing situation may result from problems in the reliability of K-Ar ages of biotite from some tuffs.

The fact that many faunal sequences do not include a Duchesnean fauna lends support to Emry's (1981) conclusion that the Duchesnean fauna of the type locality is really an early Chadronian fauna. In one area in the Vieja Group in western Texas where Duchesnean fauna is absent, an ignimbrite separates rocks containing Chadronian fauna above from those containing a Uintan fauna below. This ignimbrite has sanidine K-Ar ages ranging from 35.6 to 39.6 Ma (Wilson and others, 1968),

Table 1. Fission-track ages of zircon from tuffs in western and central Uinta basin

[Sample localities shown on plate 1. Constants: λ_f=7.03×10^{17}yr^{-1}. Number of tracks (t) counted given in parentheses. Analyst: C.W. Naeser]

			Density			Age	Number	Uranium	Location	
Loc.	Field number	Laboratory number	Fossil tracks ×10^6(t/cm^2)	Induced tracks ×10^6(t/cm^2)	Neutron dose ×10^{15}(n/cm^2)	±2σ (Ma)	of grains counted	content (ppm)	Latitude (north)	Longitude (west)
1	SRNW-83-1	DF-4783	5.63 (1120)	9.56 (952)	8.65	30.4±3.0	6	360	40°14'00"	111°11'12"
2	KL-3	DF-3351	5.37 (751)	9.86 (690)	.923	30.0±1.5	6	310	40°33'27"	110°32'56"
3	KL-2	DF-3354	2.92 (864)	4.74 (699)	.923	34.0±1.7	7	150	40°32'36"	110°32'16"
4	ICP-1A	DF-4440	5.04 (1191)	9.50 (1121)	.963	30.5±1.4	6	310	40°35'49"	109°59'52"
5	PCC-9	DF-4789	6.09 (1043)	10.04 (860)	8.53	30.9±3.1	5	370	40°30'32"	110°00'07"
6	MH-83-2	DF-4783	6.29 (1019)	10.13 (821)	8.68	32.2±3.2	6	370	40°24'27"	110°26'39"
7	BTM-4	DF-3349	5.24 (778)	9.42 (700)	.923	30.6±1.5	6	300	40°20'33"	110°32'13"
	BTM-4A	DF-4441	6.97 (1032)	11.57 (857)	.862	31.0±1.3	6	430	40°20'33"	110°32'13"
8	DM-83-1	DF-4786	6.41 (1454)	9.80 (1112)	8.60	33.6±3.1	6	360	40°26'35"	110°30'36"
9	N-83-1	DF-4778	9.56 (1593)	15.56 (1297)	8.79	32.2±2.8	6	560	40°28'45"	110°05'48"
10	N-83-2	DF-4779	5.11 (1113)	7.29 (793)	8.77	36.7±3.9	6	260	40°28'43"	110°05'54"
11	NNW-82-1	DF-4437	3.69 (444)	7.19 (433)	.936	28.7±2.0	6	240	40°25'43"	110°09'09"
12	NNW-83-1	DF-4776	4.25 (345)	6.65 (277)	8.84	33.7±5.6	3	240	40°25'43"	110°07'50"
13	NNW-83-2	DF-4777	3.43 (540)	5.49 (432)	8.82	32.9±4.5*	5	200	40°25'44"	110°09'02"
14	LA-1	DF-4438	7.36 (954)	10.77 (698)	.862	35.2±1.6	6	400	40°24'52"	109°45'43"
15	VNW-1	DF-4436	5.34 (569)	8.10 (431)	.936	36.9±1.8	6	270	40°24'33"	109°42'01"
16	FCP-26	DF-3350	6.45 (1134)	9.56 (840)	.923	37.2±1.7	6	300	40°22'40"	110°39'40"
17	FCP-27	DF-3352	5.50 (1424)	7.94 (1026)	.923	38.2±1.8	6	250	40°22'42"	110°39'57"
18	BLU-83-1	DF-4780	4.95 (916)	7.83 (725)	8.75	33.0±3.4	6	280	40°19'41"	110°09'17"
19	BLU-83-2	DF-4781	6.81 (631)	10.28 (476)	8.72	34.5±4.4	6	370	40°19'40"	110°09'21"
20	RG-1	DF-3352	6.01 (694)	8.80 (508)	.923	37.6±1.9	6	280	40°09'54"	110°33'02"
21	SRSE-3	DF-4957	3.72 (879)	3.82 (451)	.756	43.9±5.4	6	160	40°00'15"	111°05'32"
22	LC-82-1	DF-4434	4.07 (999)	5.31 (651)	.936	42.8±2.2	6	180	39°38'48"	110°37'08"
23	JH-82-1	DF-4439	6.47 (1228)	7.87 (747)	.862	42.3±2.0	6	290	39°52'57"	110°44'57"
24	WCS-11	DF-2945	2.05 (761)	6.55 (561)	1.11	44.9±2.1	6	170	40°23'15"	111°01'02"

*Age suspect because of the large number of older grains in population and the small number of countable grains.

and the mean value of these ages represents the age of the Eocene-Oligocene boundary.

Present Work

The paucity of fossils in the Duchesne River Formation precludes paleontological dating, but the presence of numerous tuff beds provides an opportunity to obtain isotopic ages. Samples were collected from 24 tuff beds, from the main body of the Green River Formation to the Starr Flat Member of the Duchesne River Formation. Sample localities are shown on plate 1, and the localities and their stratigraphic positions are described in the appendix. Zircon separates from 24 localities were dated by using the fission-track method, and biotite separates for these localities were dated by using the K-Ar method. These data are tabulated in tables 1 and 2 and shown graphically in figure 2.

In the alluvial-facies rocks of the Duchesne River Formation, relatively few tuffs contain biotite suitable for dating, whereas zircon suitable for dating is abundant. Individual zircon grains recycled from older sedimentary rocks can be differentiated from primary volcanogenic grains by their mineralogy and color. We extended our study down section in order to overlap that part of the Green River Formation dated previously by using K–Ar and ^{40}Ar^{39}Ar methods (Mauger, 1977; O'Neill and others, 1980).

Table 2. Potassium-argon ages of biotite from tuffs in the Uinta basin

[Sample localities shown on plate 1. Potassium analyses made using a lithium internal standard. Constants: ^{40}Kλ_ε=0.581×10^{-10}yr^{-1}, λ_β=4.962×10^{-10}yr^{-1}. Atomic abundances for ^{40}K/K total=1.167×10^4 mol/mol. Analysts: R.F. Marvin, H.H Mehnert, V.M. Merritt, and E.L. Brandt]

Locality	Field no.	K_2O (wt. percent)	Radiogenic ^{40}Ar (moles/g[1])	Radiogenic ^{40}Ar (percent[2])	Age ±2σ (Ma)
4	ICP-1	8.00, 8.03	4.386	83	37.6±1.4
16	FCP-26	7.11, 7.04	5.009	50	48.5±1.7
20	RG-1	7.59, 7.55	3.348	53	30.5±1.1

[1]Times10^{-10}.

[2]Of total argon.

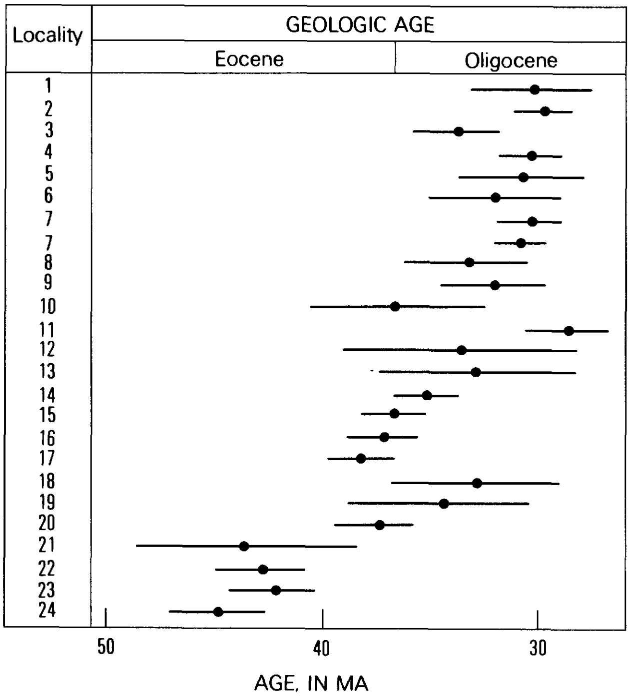

Figure 2. Fission-track ages of zircon from tuffs in upper Green River Formation (locs. 20–23), Duchesne River Formation (locs. 1–19), and conglomerate (loc. 24). Bar indicates range of analytical uncertainty. Sample localities shown on plate 1 and described in appendix.

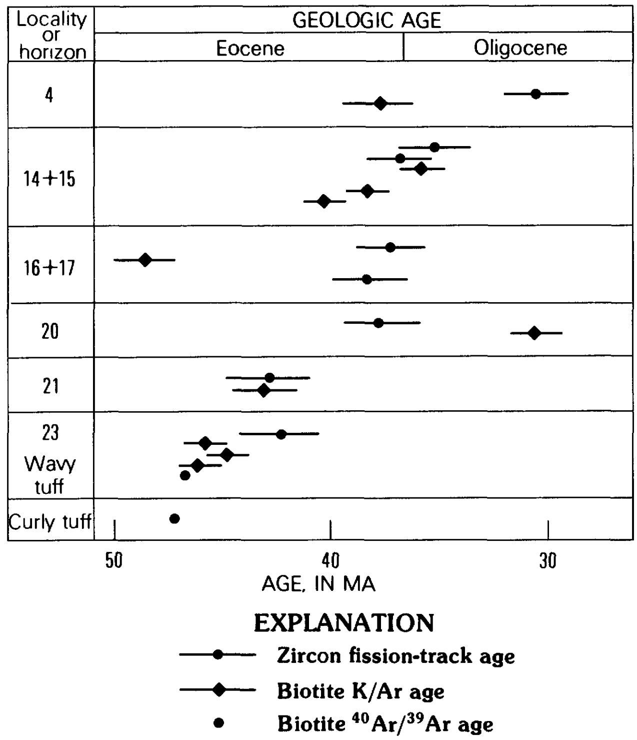

Figure 3. Comparison of biotite and zircon ages from tuffs in upper part of Green River Formation (locs. 20–23) and Duchesne River Formation (locs. 4, 14–17), Uinta basin. Bar indicates range of analytical uncertainty. Biotite $^{40}Ar/^{39}Ar$ ages for wavy tuff and curly tuff from O'Neill and others (1981.) Sample localities shown on plate 1 and described in appendix.

Analysis of radiometric data from our study revealed discrepancies between ages determined by the two methods, as well as between the determined ages and some stratigraphic relations (fig. 3). In two of the three samples for which we determined both biotite and zircon ages, the indicated ages for the biotite are from 7 (loc. 4) to 11 m.y. (loc. 16) older than those for the zircon (fig. 3). Age discrepancies such as these can result from partial annealing of the zircon, caused either by hydrothermal alteration or deep burial, that results in an indicated age that is too young; however, there is no geologic evidence to suggest that annealing occurred. Furthermore, biotite and zircon from more deeply buried tuffs in the Green River Formation show less discrepancy in age. The discordant ages could result from either excess radiogenic argon in the biotite or minor amounts of detrital biotite reworked from Precambrian rocks. Likely sources for detrital biotite in the Duchesne River Formation are the biotite-bearing shales of the Uinta Mountain Group, which crop out along the south flank of the Uinta Mountains.

At locality 20, the biotite age is 7 m.y. younger than the zircon age, at present an unexplainable discrepancy. In addition, the zircon age (37.6 Ma) is somewhat younger than the age of the sampled tuff as indicated by interfingering relations of the sandstone and limestone facies of the Green River Formation with the Uinta Formation, which contains a late Eocene mammalian fauna. It is possible that the 37.6-Ma age is correct because these beds are about 600 m above tuffs that have zircon fission-track ages of 42–43 Ma (locs. 20, 21). A compilation of subsurface data along the depositional axis of the Uinta basin confirms the relations suggested by surface mapping, shows westward thickening of lacustrine rocks at the expense of alluvial rocks (Johnson, 1985, fig. 3), and suggests that Lake Uinta persisted well into late Eocene time in the western part of the basin.

At localities 14 and 15, zircon fission-track ages were determined for a tuff that occurs with vertebrate-bearing strata of the Lapoint Member of the Duchesne River Formation. Rocks at almost the same stratigraphic horizon were previously dated by three K-Ar determinations of biotite (McDowell and others, 1973; Mauger, 1977). Both the zircon and biotite localities are near a fossil locality described by Emry (1981) and the type locality of the Duchesnean fauna. The three biotite ages are discordant, ranging from 35.7 to 40.3 Ma, and

Mauger (1977) stated that the two younger ages are probably too young. The zircon ages of 35.2 and 36.9 are in good agreement (fig. 3). The combined five ages do not conclusively date the tuff but suggest that the sediments may range in age from late Eocene to very early Oligocene. The two zircon ages and the youngest biotite age form a concordant group and suggest a very early Oligocene age, an age that agrees with that of nearby rocks containing earliest Chadronian fauna (Emry, 1981).

The K-Ar ages of biotite and fission-track ages of zircons for tuffs in the Green River Formation vary in their agreement with each other (fig. 3). In the sandstone and limestone facies near the top of the Green River Formation (loc. 20), the zircon age is considerably older (discussed above) than the age for coexisting biotite. Four biotite ages (Mauger, 1977; O'Neill and others, 1981) from the Wavy tuff in the upper part of the main body of the Green River Formation are in approximate agreement, at about 46 Ma; a zircon age from the same bed (loc. 23) is younger (42.8 Ma). Biotite ages (Mauger, 1977) and zircon ages from our study for a tuff just below the base of the saline facies in Indian Canyon (loc. 22) agree quite well at 42–43 Ma. A zircon age for a tuff from the lower part of the saline facies south of Strawberry Reservoir (loc. 21) is in the same time span.

In the eastern part of the Salt Lake City $1° \times 2°$ quadrangle, a series of tuff samples (locs. 9–13, 18, 19) was collected from a stratigraphic sequence that extends from the upper part of the main body of the Green River Formation through the Duchesne Formation and includes the type sections of the Dry Gulch Creek and Starr Flat Members of the Duchesne River Formation (Andersen and Picard, 1972). A plot of the stratigraphic position of these samples and their zircon fission-track ages is shown in figure 4. The zircon fission-track ages agree in gross aspect with the order of superposition but have a number of discrepancies. The age for locality 11 appears to be too young and the age for locality 10 too old; however, the range of analytical uncertainty for each sample is sufficiently large such that the error bar almost intersects the best-fit line through the zircon ages. This plot indicates an age of about 33 Ma for the base of the Lapoint Member, an age that is 1–3 m.y. younger than the ages of 35 and 37 Ma determined for the basal beds at localities 14 and 15, some 40 km to the east. This east to west younging of the basal part of the Lapoint Member is indicated by a zircon age of 32 Ma for locality 6, some 25 km farther west, about as far west as the Lapoint Member has been identified (fig. 5).

Zircons from tuffs in the coarse-grained proximal facies of the Starr Flat Member were dated at localities 2, 3 and 4 and from a tuff in an intermediate facies at locality 5. The proximal facies generally is poorly exposed because it is poorly cemented and is at an altitude where soils, vegetation, and colluvial cover are all well developed. About 4 km southeast of Moon Lake, however, conglomerate and sandstone in some excellent exposures dip gently northward toward the Uinta Mountains as a result of local renewed movement on a part of the South Flank fault zone. Zircons from two tuff beds (locs. 2, 3) gave ages of 30 and 34 Ma, respectively, with no overlap of analytical uncertainties (fig. 3). The older zircon is from a tuff that is only about 60 m stratigraphically lower than the tuff containing the younger zircon, and the 4-m.y.-age difference seems excessive. The age difference could be real, however, because the sample locality is at the margin of the basin and numerous interruptions of sedimentation may have occurred. The third age determination, 30.5 Ma, is for a zircon from a tuff bed in the proximal facies of the Starr Flat Member (loc. 4) in an area where fine-grained rocks to the east, mapped as Lapoint Member, interfinger with coarser grained rocks to the west, mapped as Starr Flat Member. The tuff bed is overlain by a poorly cemented cobble and boulder conglomerate that caps the flat-topped shoulder of the Uinta Mountains to the north and has been called Bishop Conglomerate.

A few isotopic ages have been determined for rocks from the Bishop Conglomerate. Concordant K-Ar ages of hornblende and biotite from a tuff in the Bishop Conglomerate about 60 km east of locality 4 are 28.6 and 29.5 Ma, respectively, (Hansen and others, 1981). A K-Ar age for biotite from a tuff about 1 km away and at a similar stratigraphic horizon is 26.9 Ma, whereas the age for biotite from a tuff in the Bishop Conglomerate about 30 km to the southeast is 42.4 Ma (Winkler, 1970). The latter two ages are questionable because no other mineral in the tuff was dated. The age of 26.9 Ma may be correct, but the age of 42 Ma probably is too old. In view of the ages presently available, we conclude that the part of the Starr Flat Member of the Duchesne River Formation from which we obtained ages of 30–31 Ma (locs. 2, 4, 5) may be a time equivalent to at least part of the Bishop Conglomerate.

As previously mentioned, the conglomerate mapped as Starr Flat Member at the northern edge of the Uinta basin at Dry Mountain and Tabby Mountain grades southward into the basin to sandstone mapped as undivided Duchesne River Formation. South of Dry Mountain, near the structural axis of the basin and on trend with the conglomerate of the Starr Flat Member that forms Dry Mountain, two samples of zircon from a tuff in rock mapped as undivided Duchesne River (loc. 7) gave ages of 31.0 and 30.6 Ma, similar to those obtained from the proximal facies of the Starr Flat Member (locs. 2–5). These ages at locality 7 extend the westward-younging trend of rocks of the Duchesne River Formation along the axis of the Uinta basin (fig. 5).

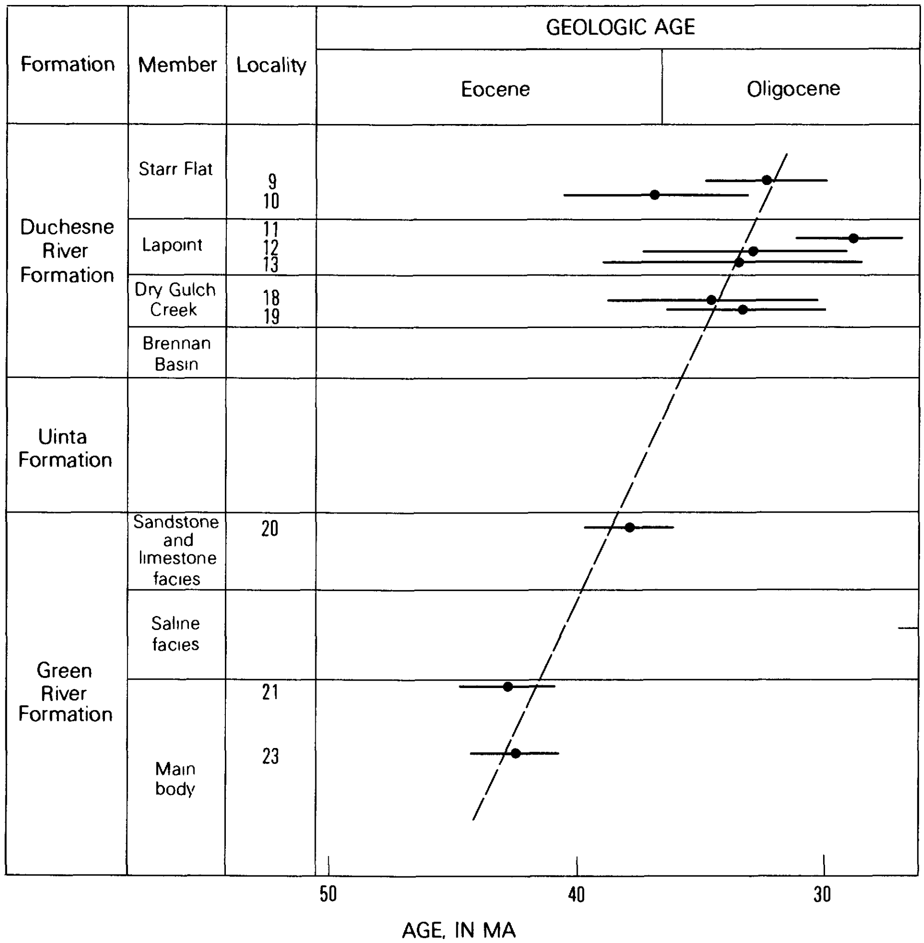

Figure 4. Fission-track ages of zircon from tuffs in upper part of Green River Formation (locs. 20, 21, 23) and in Duchesne River Formation (locs. 9–13, 18, 19) in pseudosection in Indian Canyon-Bluebell-Neola-Uinta River area, including the type area of the Dry Creek Gulch and Starr Flat Members of the Duchesne River Formation. Bar indicates range of analytical uncertainty; dashed line is best fit as determined visually. Sample localities shown on plate 1 and described in appendix.

An obvious angular unconformity between undivided Duchesne River Formation and the Starr Flat Member exists at the margin of the basin in the Dry Mountain area. The approximate age of this unconformity and the length of the hiatus it represents can be inferred from ages determined in our study. Tuffs within the undivided Duchesne River yield ages of 37.2 and 38.2 Ma (locs. 16, 17), whereas tuffs within the overlying Starr Flat Member yield ages of 30.0 and 34.0 Ma. These data suggest an age of 37–34 Ma and a duration of hiatus of 3 to 4 m.y.; however, the zircon age of 33.6 Ma from a tuff (loc. 8) in the rocks below the unconformity east of Rock Creek is intermediate between the two pairs just mentioned. The analytical uncertainty of that age is about twice that of the ages from localities 16, 17, 2, and 3, and, because we have only one age determination at locality 8, perhaps the discrepancy between that age and those for localities 16 and 17 should not be emphasized.

To the west, in the graben north of Strawberry Reservoir, zircons from a clay bed, an altered tuff, in rocks mapped as undivided Duchesne River Formation gave a fission-track age of 30.4 Ma (loc. 1). In the graben, volcanic rock clasts in some of the conglomerates indicate that a volcanic field was one of the source areas for the sediments. The nearest remnant of a volcanic field is 15 km to the north, and a sample of a lava flow from this field gave a K-Ar age for biotite of 34.9 Ma (Crittenden and others, 1973). This remnant is the most southerly outlier of the Keetley volcanic field. Stratigraphic relations between the dated tuff and the volcanic-bearing conglomerate north of Strawberry Reservoir are not exposed, but the two rock types are believed to be in about the same part of the section.

Zircon from a tuff bed in a conglomerate (map unit Tc, plate 1) underlying the Keetley Volcanics just north of the margin of the Uinta basin (loc. 24) gave a

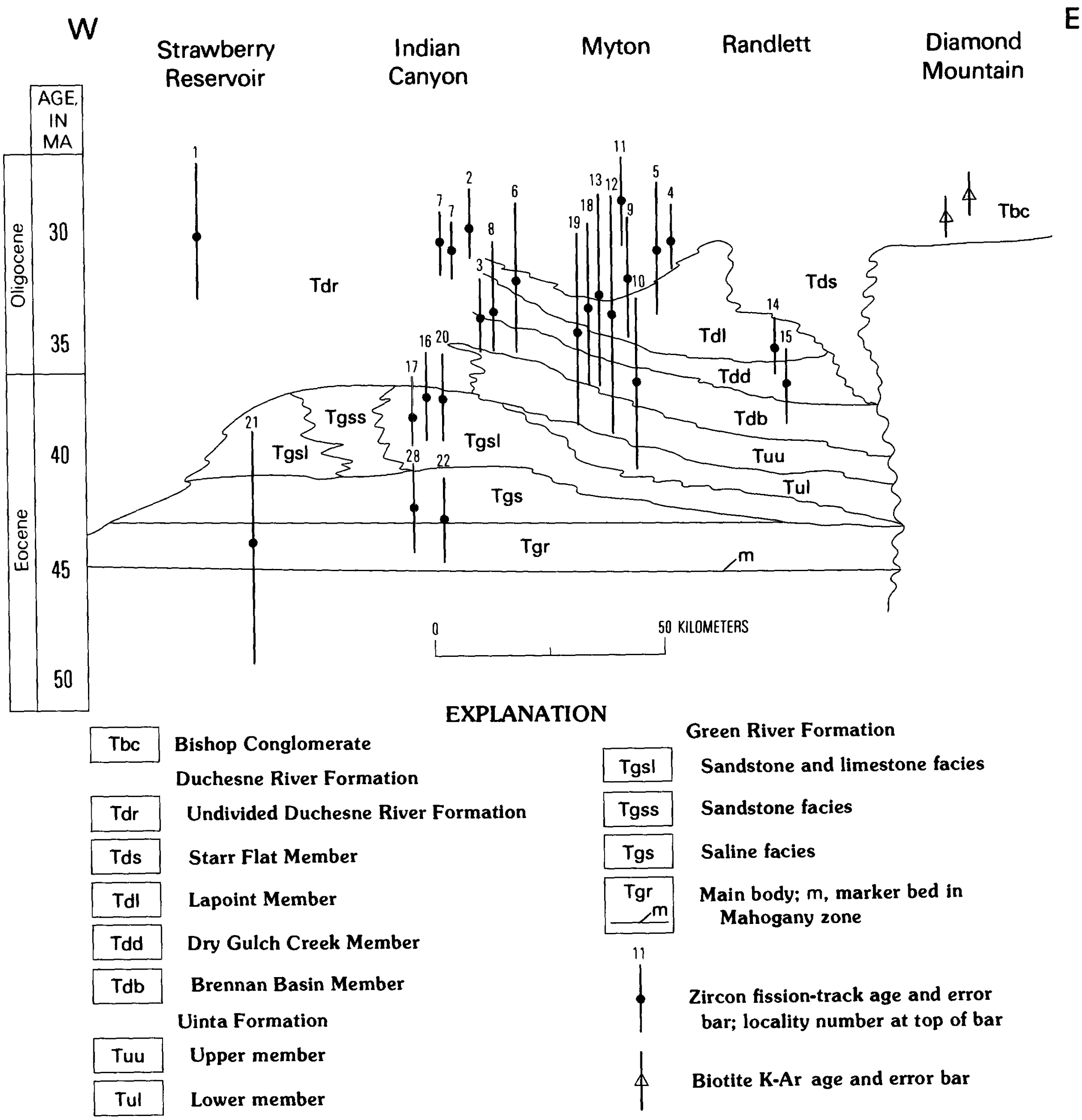

Figure 5. Age and distribution of upper part of Green River Formation, Uinta Formation, Duchesne River Formation, and Bishop Conglomerate in east-west cross section along Uinta basin projected to a vertical plane. North-south changes are not shown well, but sample localities span those changes. Fission-track ages shown in table 1; sample localities shown on plate 1. K-Ar argon ages of Bishop Conglomerate from Winkler (1970) and Hansen and others (1981).

fission-track age of 44.9 Ma. This age is older than anticipated because in many places the conglomerate appears to grade into volcanic-rock-bearing conglomerate that must be younger than 38 Ma, the age of the oldest Tertiary igneous rocks dated nearby. Locality 24 is about 4 km from the nearest occurrence of volcanic clasts and probably at least 100 m stratigraphically below that occurrence. At locality 24, the conglomerate is 100–150 m thick and composed of Paleozoic carbonate rocks and sandstone and a small proportion of Precambrian quartzite. It is poorly sorted and crudely bedded in units as thick as 5 m and contains angular to subrounded boulders as much as 3 m in diameter. The large clast size and poorly sorted nature of the deposit indicate that the source of sediments was high land in the western part of the Uinta Mountains. The zircon age of the tuff indicates that the tuff was deposited at the same time as the upper part of the main body of the Green River

Formation (fig. 2). The conglomerate is the same age as a conglomerate on the northern side of the Uinta Mountains that interfingers with the Bridger and Green River Formations of the Green River basin (Bradley, 1964) (unpublished mapping by Bryant and fission-track zircon ages by Naeser).

The results of a few K-Ar or fission-track age determinations of minerals do not furnish sufficient data with which to decipher the complete stratigraphic and structural history of a region. Ideally, age determinations by more than one method on a number of samples should be done. Spurious ages are more easily recognized, and the greater number of ages provide additional constraints to geochronological interpretations. Biotite separated from a tuff may have been contaminated with biotite from older rocks or potassium may have been leached from the biotite during weathering. Because fission-track ages from zircon have greater analytical uncertainty than K-Ar ages from biotite, correlations between tuffs dated only by using fission-track ages from zircon are less certain. In the samples from the Uinta basin, the analytical uncertainty of zircon ages ranges from 1.3 to 5.6 m.y.

Source of the Tuffs

Tuffs in the Duchesne River Formation range in age from 30 to 38 Ma. To the north and west, several volcanic formations of similar age crop out. K-Ar ages determined for the Keetley Volcanics north and west of the Uinta basin range from at least 34 to 38 Ma (Crittenden and others, 1973; Bromfield and others, 1977). K-Ar ages for volcanic rocks of the Moroni and Goldens Ranch Formations, which unconformably overlie the southwestern extension of the Uinta basin, range from 30 to 39 Ma (Witkind and Marvin, 1989).

The sources of the volcanic rocks and the tuffs in the Duchesne River Formation were probably associated intrusive rocks. Ages of intrusive rocks in the Wasatch Range, just west of area of the Keetley Volcanics, indicate a much longer age span for igneous activity. Biotite K-Ar ages range from 25 to 38 Ma, hornblende ages from 36 to 47 Ma, zircon fission-track ages from 24 to 43 Ma, and sphene fission-track ages from 23 to 36 Ma (Crittenden and others, 1973; Bromfield and others, 1977). Hornblende ages are older than biotite ages by varying amounts; the only concordant pair gives an age of about 34 Ma. Another sample from the same pluton gives the most discordant biotite-hornblende pair. The intrusive rocks in the Wasatch Range were probably emplaced from about 26 to 37 Ma and may be the roots of volcanoes that erupted the tuffs deposited in the Uinta basin, 60–170 km to the east. In addition to potential source volcanoes in the Wasatch Mountains, two other possible sources for tuffs are the relatively nearby Oquirrh Mountains and East Tintic Mountains, about 90 km northwest and west of the Uinta basin, respectively (fig. 1). Ages of igneous rocks in the Oquirrh Mountains range from 33 to 40 Ma (Moore and others, 1968), but the most intense activity occurred from 37 to 40 Ma (Warnaars and others, 1978). In the East Tintic Mountains, intense igneous activity was from 32 to 33 Ma, although sporadic activity occurred from 16 to possibly 40 Ma (Laughlin and others, 1969; Morris and Lovering, 1979).

Andersen and Picard (1974) found that strontium, zirconium, and rubidium contents of altered tuffs in the Duchesne River Formation are similar to those of samples of the Keetley Volcanics but dissimilar to samples of the Absaroka Volcanic Supergroup and Challis Volcanics, two other potential sources of tuff during the Paleogene. The Absaroka and Challis rocks are 400 km north and 500 km northwest of the Uinta basin, respectively, and older than the Duchesne River Formation; however, they are potential sources for tuffs in the Green River Formation.

FUTURE STUDIES

Paleomagnetic methods could be used to correlate and date deposits of the Uinta basin. The Eocene-Oligocene time line possibly could be located and traced throughout the basin by locating Chron C 13 of the magnetic time scale in the rocks. Chron C 13 is the interval in which the Eocene-Oligocene time boundary is now placed (Berggren and others, 1985; Prothero, 1985), and our localities 14 and 15 are probably in that chron. If our fission-track ages are correct (excluding obviously questionable ages), then deposition of the Duchesne River Formation occurred during the early Oligocene in Chrons C 10, C 11, C 12, and C 13, and perhaps as early as Chrons C 15 and C 16.

Another possible approach for improving our understanding of the time-stratigraphic framework of the Duchesne River Formation is by correlating tuff beds based on studies of texture, mineralogy, chemistry, and stratigraphy. Deposits of Quaternary gravel conceal large areas in the central part of the basin, however, and prevent tracing of outcrops of individual beds.

CONCLUSIONS

The isotopic ages of from 30 to 38 Ma of zircon from tuffs in the Duchesne River Formation indicate that the upper three members are of Oligocene age and that rocks at the type locality of the Duchesnean fauna are of earliest Oligocene age. The basal Brennan Basin Member in the eastern part of the basin is late Eocene in age, as indicated by Uintan fauna collected by previous investigators. The upper part of the Brennan Basin Member in the central part of the basin may be early Oligocene in age.

The mapping and isotopic ages presented in this report support the conclusions of Andersen and Picard (1972, 1974) that the members of the Duchesne River Formation members, as they defined them, are time transgressive and that the tuff beds of the Duchesne River Formation probably came from volcanoes in the Wasatch Range, East Tintic Mountains, and Oquirrh Mountains to the west. The lower part of the Duchesne River Formation in the western part of the basin is older than the basal part in the eastern part of the basin. To the east, near Duchesne, the lower part intertongues with the upper part of the Uinta Formation of late Eocene age, whereas, to the west, it intertongues with progressively older units including the sandstone and limestone facies and the saline facies of the Green River Formation and the main body of the Green River Formation of middle Eocene to late Paleocene age near the margin of the area of deposition of lacustrine rocks.

The unconformity below the conglomerates of the Starr Flat Member along the northern margin of the basin indicates that significant tilting occurred in early Oligocene time as a result of uplift of the Uinta Mountains or downwarping of the Uinta basin. This unconformity is well developed in the Little Mountain area, where conglomerate and sandstone of the Brennan Basin Member dip 15°-30° to the south and southwest and are overlapped by very gently dipping conglomerate and sandstone of the Starr Flat Member. On Little Mountain, the Starr Flat Member grades upward into what was mapped as Bishop Conglomerate in the Vernal 1° × 2° quadrangle (Rowley and others, 1985).

The Duchesne River Formation in the central and eastern Uinta basin was deposited in latest Eocene and early to middle Oligocene time. The Starr Flat Member lacks a well-defined contact with the overlying Bishop Conglomerate. Although much of the Uinta Mountain area was stable during deposition of the Bishop Conglomerate, the Uinta basin was still sinking relative to the Uinta Mountains; thus, while pediment gravels were being deposited on the northern flank of the Uinta Mountains and across the Uinta Mountains east of the Uinta basin, a thick alluvial-fan deposit was prograding into the basin. At the margin of the basin, facies changed from dominantly conglomerate to dominantly sandstone in a distance of only 10 km in 30-m.y.-old rocks of the Starr Flat Member of the Duchesne River Formation. The gentle fold that forms the axis of the Uinta basin at the surface developed after deposition of the Starr Flat Member, and the South Flank fault was locally reactivated. This reactivation movement may have been a weak western manifestation of Neogene faulting in the eastern part of the Uinta uplift (Hansen, 1984).

As Lake Uinta shrank, detritus from relatively distant sources to the east and southeast was deposited to form the Uinta Formation in the east-central and eastern parts of the Uinta basin. In the west-central part of the basin, west of Duchesne, alluvial deposits overlying the Green River Formation were derived from the Uinta uplift to the north. In the east-central and eastern parts of the basin, the onset of deposition of the Duchesne River Formation, which was derived from relatively nearby sources in the Uinta Mountains to the north, signaled the beginning of an episode of uplift of the southern part of the Uinta Mountains during latest Eocene and Oligocene time. No evidence of uplift of the Uinta Mountains during this time has been found in the Green River basin on the northern side of the Uinta Mountains. In the east-central and eastern Uinta basin, differences in source area and distance of transport of sediments of the Duchesne River and Uinta Formations account for their distinct characteristics. In the west-central part of the basin, no such contrast of source and distance of transport exists for alluvial rocks overlying the lacustrine rocks of the Green River Formation, and all alluvial rocks have Duchesne River lithologies. Consequently, all of the Paleogene alluvial-facies rocks overlying the Green River Formation in the western part of the basin are included in the Duchesne River Formation, even though we know that some of them must be chronostratigraphic equivalents of the Uinta Formation.

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Appendix. Description of sample localities
[Localities shown by number on plate 1]

1. Tuff from cut bank in Clyde Creek, 360 m S. 75° W. from junction of Clyde Creek and Strawberry River. Undivided Duchesne River Formation in the Strawberry graben. 170 m S. 62° E. of center of sec. 12, T. 35 N., R. 12 W., Wasatch County. Strawberry Reservoir Northwest quadrangle.

2. Lens of tuffaceous sandy limestone in conglomerate and subordinate sandstone, about 200 m thick, on south flank of Uinta Mountains. Starr Flat Member of the Duchesne River Formation. East face of Dry Ridge, 10,640 ft altitude. 340 m N. 11° E. of center of sec. 22, T. 2 N., R. 6 W., Duchesne County. Kidney Lake quadrangle.

3. Tuff at base of very pale greenish yellow weathering clay, 0.5-1 m thick, in predominantly conglomeratic section about 200 m thick on south flank of Uinta Mountains. Starr Flat Member of the Duchesne River Formation. East face of Dry Ridge, 10,590 ft altitude. 180 m N. 45° E. of spot elevation 10,811; about 60 m stratigraphically below sample locality 2; sec. 26, T. 2 N., R. 6 W., Duchesne County. Kidney Lake quadrangle.

4. Brownish-gray biotitic clay from near base of tuff, 4-5 m thick and 3.5 m below sample of tuff from which biotite was dated. Underlies boulder conglomerate called Bishop on some maps; overlies sandstone and claystone of Duchesne River Formation; here included in Starr Flat Member. Below outward bend on Forest Service Road, 9,060 ft altitude. Sec. 4, T. 2 N., R. 1 W., Duchesne County. Ice Cave Peak quadrangle.

5. Clay containing some megascopic biotite grains. Starr Flat Member of Duchesne River Formation. To the east, rocks become finer grained and have been mapped as Lapoint Member by Rowley and others (1985). From borrow pit along road between Whiterocks and the Uinta River, 170 m west of eastern boundary of Pole Creek Cave quadrangle. Center of south half of sec. 4, T. 1 N., R. 1 W., Duchesne County.

6. Greenish-gray to gray clay containing some fine-grained biotite. Base of the Lapoint Member of Duchesne River Formation. Roadcut on road between Mountain Home and Rock Creek, 60 m SE of spot elevation 7090. NW1/4SW1/4 sec. 10, T. 1 S., R. 5 W., Duchesne County. Mountain Home quadrangle.

7. Greenish-gray to dark-gray biotite tuff, at least 1 m thick. Underlies thick-bedded, yellowish-gray sandstone containing a few light-red beds. Undivided Duchesne River Formation. Roadcut on road to drill site. 190 m from west line, 30 m from south line, sec. 35, T. 1 S., R. 6. W., Duchesne County. Blacktail Mountain quadrangle. Two samples collected.

8. Light-gray clay containing fine-grained biotite. East side of gulley, 7,430 ft altitude; 1 m above gray clay-rich interval. Near base of Tertiary rocks close to the margin of the basin. Undivided Duchesne River Formation. 480 m S. 44° W. from spot elevation 7885 on north line of sec. 36, T. 1 N., R. 6 W., Duchesne County. Dry Mountain quadrangle.

9. Greenish-gray biotite clay. Probably unit 7 of type section of Starr Flat Member of Duchesne River Formation (Andersen and Picard, 1972). Highest tuff in good exposures south of John Starr Flat near base of 10-m-thick sequence of tuffaceous rock and 150 m above base of Starr Flat Member. 400 m S. 43° E. of triangulation point Neola 7226, 6,950 ft altitude. Center of south half of sec. 15, T. 1 N., R. 2 W., Duchesne County. Neola quadrangle.

10. Greenish-gray clay containing very minor amounts of fine-grained biotite and possible some silt. In 18-m-thick interval of tuffaceous material, about 7 m below top and about 25 m below sample 9. Type section of Starr Flat Member of Duchesne River Formation. SE1/4SW1/4 sec. 15, T. 1 N., R., 2 W. Duchesne County, Neola quadrangle.

11. Biotitic clay in cut made by Yellowstone Feeder canal on east side of Monarch Ridge, at 6,620 ft altitude. About 150 m above base of Lapoint Member of Duchesne River Formation. Sec. 6, T. 1 S., R. 2 W., Duchesne County, Neola Northwest quadrangle.

12. Greenish-gray altered tuff from open area west of new reservoir. About 10 m above base of Lapoint Member of Duchesne River Formation. About 6,150 ft altitude, 480 m from north line and 1,020 m from west line of sec. 5, T. 1 S., R. 2 W., Duchesne County. Neola Northwest quadrangle.

13. Gray silty clay in gulley made by Yellowstone Feeder canal. About 60 m above sample locality 12 and 50 m below sample locality 11. Lapoint Member of Duchesne River Formation. NE1/4 sec. 6, T. 1 S., R. 2 W., Duchesne County. Neola Northwest quadrangle.

14. Gray biotitic clay, 0.5 m thick. Base of the Lapoint Member of Duchesne River Formation. From roadcut on Utah Highway 121 about 100 m north of BM 5769. Sec. 2, T. 5 S., R. 19 E., Uintah County. Lapoint quadrangle.

15. Biotitic clay. Base of Lapoint Member of Duchesne River Formation. From cut just off Utah Highway 121, 270 m S. 32° W. of center of sec. 5, T. 5 S., R. 20 E., Uintah County. Vernal Northwest quadrangle.

16. Olive-gray biotitic clay, 0.5 m thick, overlying bluish-white to yellowish-gray sandstone. Thin basin margin section, undivided Duchesne River Formation; 880 m above Currant Creek Formation. 7,320 ft altitude, 110 m S. 21° W. of center of sec. 22, T. 1 S., R. 7 W., Duchesne County. Farm Creek Peak quadrangle.

Appendix. Continued

17. Biotitic tuff, 0.3 m thick, underlain by tuffaceous sandstone and overlain by tuffaceous siltstone in section containing light-gray and moderate-reddish-brown sandstone and siltstone; 80 m stratigraphically below sample locality 16. Undivided Duchesne River Formation. 7,100 ft altitude on south side of gulley, 430 m S. 83° W. of center of sec. 22, T. 1 S., R. 7 W., Duchesne County. Farm Creek Peak quadrangle.

18. Gray clay, containing some fine-grained biotite, in a clay sequence, 1-m-thick that is below a yellowish-gray sandstone, 1-2 m thick. About 70 m above base of Dry Creek Gulch Member in type section of Duchesne River Formation, probably unit 3 of Andersen and Picard (1972). Altitude 6,070 ft, 380 m N. 13° E. of Ioka 6268 triangulation point, NE1/4NW1/4 sec. 7, T. 2 S., R. 2 W., Duchesne County. Bluebell quadrangle.

19. Light-olive-green silty clay. Type section of Dry Creek Gulch Member of Duchesne River Formation. 330 m N. 4° W of Ioka 6268 triangulation point; about 10 m above sample locality 18. NW1/4NW1/4 sec. 7, T. 2 S., R. 2 W., Duchesne County. Bluebell quadrangle.

20. Pale-yellowish-orange-weathering white biotitic tuff, 0.6-0.8 m thick, in thickly bedded shale and siltstone. About 30 m above base of sandstone and limestone facies of Green River Formation. Roadcut on new road on south side of Strawberry River valley, 870 m west of sharp bend in old road along Strawberry River. Sec. 3, T. 4 S., R. 6 W., Duchesne County. Rabbit Gulch quadrangle.

21. Tuff, about 0.5 m thick, about 70 m above base of saline facies of Green River Formation. On ridge, 8,680 ft altitude, SE corner of sec. 26, T. 6 S., R. 7 E., Wasatch County. Strawberry Reservoir Southeast quadrangle.

22. Biotitic tuff, about 10 m below base of saline facies of Green River Formation. Age of biotite from this locality of 42.9 Ma (recalculated) was considered too young by Mauger (1977). 70 m N. 20° W. of Gulf Oil Indian Canyon No. 1 well (drill-hole symbol shown on topographic map), sec. 12, T. 6 S., R. 7 W., Duchesne County. Lance Canyon quadrangle.

23. Biotitic wavy tuff. Upper part of main body of Green River Formation above Mahogany oil shale zone. Recalculated potassium-argon age of biotite from this locality is 46.0 Ma (Mauger, 1977). From roadcut on Utah Highway 33, 140 m along road from west edge of Jones Hollow quadrangle. Sec. 12, T. 11 S., R. 10 E., Duchesne County.

24. Greenish-white clay, about 35 m thick, in yellowish-brown to brown silty clay in a sequence of conglomerate and sandstone of map unit Tc (plate 1). Steep slope north of "t" in White Ledge, altitude 9,960 ft. About 45 m above base of conglomerate; overlain by Keetley Volcanics 5 km to the northeast. Sec. 16, T. 1 S., R. 10 W., Wasatch County. Wolf Creek Summit quadrangle.

Chapter K

Ages of Late Paleogene and Neogene Tuffs and the Beginning of Rapid Regional Extension, Eastern Boundary of the Basin and Range Province near Salt Lake City, Utah

By BRUCE BRYANT, C. W. NAESER, R. F. MARVIN, and H. H. MEHNERT

A multidisciplinary approach to research studies of sedimentary rocks and their constituents and the evolution of sedimentary basins, both ancient and modern

U.S. GEOLOGICAL SURVEY BULLETIN 1787

EVOLUTION OF SEDIMENTARY BASINS—UINTA AND PICEANCE BASINS

CONTENTS

PLATE

[Plate is in pocket]

FIGURES

TABLES

CONVERSION FACTORS FOR SOME SI METRIC AND U.S. UNITS OF MEASURE

To convert from	To	Multiply by
Feet (ft)	Meters (m)	0.3048
Miles (mi)	Kilometers (km)	1.609
Pounds (lb)	Kilograms (kg)	0.4536
Degrees Fahrenheit (°F)	Degrees Celsius (°C)	Temp °C=(temp °F–32)/1.8

Ages of Late Paleogene and Neogene Tuffs and the Beginning of Rapid Regional Extension, Eastern Boundary of the Basin and Range Province near Salt Lake City, Utah

By Bruce Bryant, C.W. Naeser, R.F. Marvin, *and* H.H. Mehnert

Abstract

Several lines of evidence, including the distribution of major rock sequences and radiometric ages from tuffs, indicate that the pattern of deposition near Salt Lake City changed in late Tertiary time. An older depositional sequence, consisting of widespread volcanic flows and tuffs, conglomerates, and minor limestone, is preserved in Neogene structural lows that flank the Wasatch fault zone. The contained volcanic rocks have ages of from 27 to 39 Ma. A younger depositional sequence, consisting of fanglomerates, volcanic tuffs, and limestones, is preserved in Neogene basins that are chiefly within the Basin and Range province. The volcanic rocks in this younger sequence range in age from 4 to 10 Ma. Sedimentation rates based on data derived from a hole drilled in a graben beneath Great Salt Lake suggest that the graben formed about 12 Ma. This age implies that rapid regional extension was probably underway 10–12 Ma, although earlier movement occurred on some Basin and Range faults.

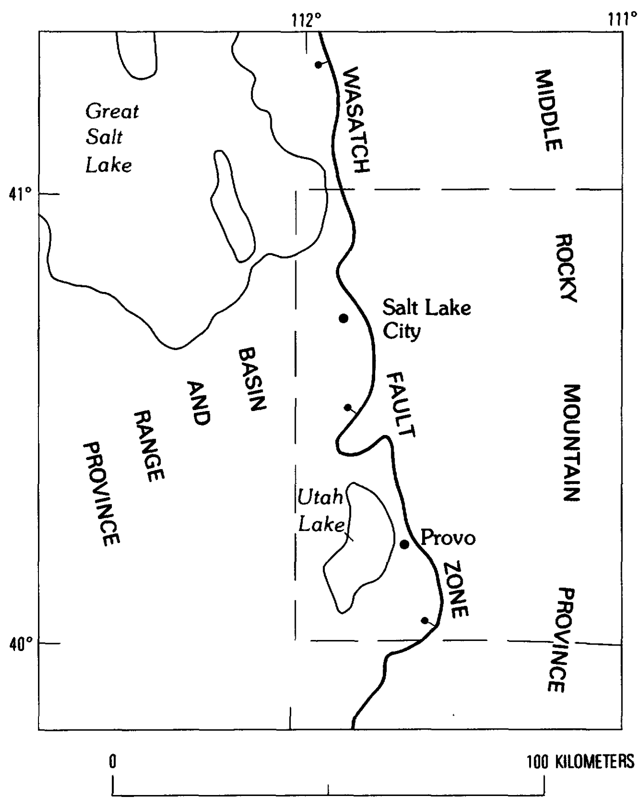

Figure 1. Boundary between Basin and Range and Middle Rocky Mountains provinces and location of Wasatch fault zone (bar and ball on downthrown side). Dashed line encloses area of western part of Salt Lake City 1°×2° quadrangle. Area of figure is the same as that of plate 1.

INTRODUCTION

Tertiary rocks near the eastern edge of the Basin and Range province (fig. 1), in the area of Salt Lake City and Provo, Utah, reflect the late Paleogene and Neogene structural history of the region, especially the timing of the inception of regional extension. In an attempt to better understand this structural history, we studied the stratigraphic relations and determined ages of some of the Tertiary rocks. In this report, we integrate these new data with previous published material.

Two rock sequences apparently bracket the time of development of modern Basin and Range topography. An older sequence, composed of volcanic and volcaniclastic rocks of predominantly latitic composition and associated sedimentary rocks, blanketed much of the region during late Eocene and Oligocene time. A younger sequence, composed of fanglomerate, limestone, and tuff, formed during Neogene time and is confined to basins that developed at that time.

Rock units in the older sequence include the Keetley Volcanics and the Traverse Volcanics of Slentz (1955), the Norwood Tuff, the Moroni, Tibble, and Goldens Ranch Formations, and a number of other units exposed in the East Tintic Mountains (plate 1). Most of these volcanic rocks are characterized by phenocrysts of plagioclase, biotite, hornblende, and, less commonly, pyroxene; consequently, they are referred to as andesite in much of the literature. Rock units in the younger sequence are grouped and collectively called the Salt Lake Formation.

The beginning of rapid regional extension in this part of the Basin and Range province probably marks the first stages in the development of modern Basin and Range topography. If so, then rocks of the Salt Lake Formation should record the inception of that movement.

Previous studies of the Tertiary history of the region (Eardley, 1955; Heylmun, 1965) were handicapped by a lack of age control: diagnostic fossils are few and difficult to find, and radiometric dates were sparse. As a result, rocks were correlated by using lithologic similarities, a reasonable approach under the circumstances. The distribution and known ages of Tertiary strata led Eardley (1955) to suggest that Basin and Range deformation began prior to the eruption of middle Tertiary rocks, rocks now known to be late Eocene and Oligocene in age.

MIDDLE TERTIARY VOLCANIC ROCKS

Volcanic rocks of intermediate composition are preserved throughout the area (plate 1), but as yet no regional stratigraphic study of them has been published. Although most of these rocks have been called andesite, chemical analyses (Moore, 1973; Bromfield and others, 1977; Morris and Lovering, 1979) show that they are latite, quartz latite, and rhyodacite. These extrusive rocks are cut by intrusions of similar composition. Other intrusions, cutting older rocks in the present-day mountain ranges, are probably the source for some of the volcanic rocks.

The volcanic rocks consist of flows, flow breccias, lahars, and tuffs and are interbedded with sedimentary rocks derived from them. The largest area of volcanic rocks, the Keetley volcanic field, occupies a structural low between the Wasatch Mountains and the west end of the Uinta Mountains (plate 1). Intrusions presumed to be feeders for the Keetley Volcanics are aligned along and west of the trend of the Uinta Mountains; some intrusions cut volcanic units, and others are exposed farther west in the uplifted block of the Wasatch Range. North and south of the axis of the volcanic field, as marked by the aligned intrusions, sequences of predominantly flows and lahars grade laterally into sequences richer in tuff, tuffaceous sediments, and conglomerate containing clasts of volcanic rock. At its base, the Keetley Volcanics overlie and interfinger with conglomerate composed of sedimentary detritus including distinctive clasts from the Jurassic(?) and Triassic(?) Nugget Sandstone. The Nugget Sandstone is a significant part of the substrate of the Keetley volcanic field as indicated by (1) the areal distribution of formations adjacent to the field, (2) the distribution of formations in windows through the field, and (3) the distribution and lithology of exotic blocks in the Keetley Volcanics (Bromfield and others, 1970; Bromfield and Crittenden, 1971). Northwest of the Keetley Volcanics, a transitional facies of the volcanic sequence is preserved in the East Canyon graben, and a distal facies is preserved in Morgan Valley (plate 1). Both of these basins were downfaulted in Neogene time. Rocks in the East Canyon graben are volcaniclastic sandstones and conglomerates, tuffs, and, less commonly, lahars, flow breccias, and ash flows; together they attain a maximum thickness of 1,000 m. Rocks in the Morgan Valley are tuffs, tuffaceous sandstones, and conglomerates and are designated the Norwood Tuff. Another area of distal-facies volcaniclastic rocks related to the Keetley Volcanics is north of Chalk Creek near the northeast corner of the map area, where a cobble-and-boulder conglomerate rich in andesitic clasts is preserved in a half graben.

West of the Keetley volcanic field and near the crest of the Wasatch Range, sedimentary and volcanic conglomerates and tuffs called the Tibble Formation (Baker and Crittenden, 1961; Baker, 1964) are preserved. No paleocurrent data have been reported from the Tibble, but we assume that it is derived from the nearby Keetley volcanic field, although it could be derived from the volcanic rocks in the Traverse Mountains to the west. The Tibble lacks any detritus from presently exposed intrusions such as the Little Cottonwood batholith or the Alta stock, both of which are closer than either the Keetley Volcanics or the Traverse Volcanics of Slentz (1955) but were not unroofed at the time the Tibble was deposited. Farther south, near and south of the Spanish Fork River, tuffs and conglomerates rich in volcanic clasts have been correlated with the Moroni Formation (Witkind and Weiss, 1985; Witkind and Marvin, 1989). The source of these volcanic sediments may have been to the west (Witkind and Marvin, 1989) in the Tintic Mountains, which contain a large volcanic field and an associated

Table 1. Previously reported ages of volcanic rocks from the western part of the Salt Lake City 1° × 2° quadrangle and vicinity

[Sample localities shown on plate 1. Converted to IUGS constants (Dalrymple, 1979, table 2). Method: F-t, fission track; K-Ar, potassium-argon]

Locality	Mineral or whole rock	Method	Age ±2σ (Ma)	Unit name or type of deposit	Reference
4	Zircon	F-t	10.3±1.0	Salt Lake City Formation	Naeser and other(1983)
11	Zircon	F-t	6.5±0.5	Salt Lake City formation	Naeser and others (1983)
12	Biotite	K-Ar	45.0±1.4	Tuffaceous deposits	Van Horn (1981)
12	Zircon	F-t	37.4±1.6	Tuffaceous deposits	Van Horn (1981)
12	Biotite	K-Ar	37.7±1.1	Volcanic breccia	Van Horn (1981)
12	Zircon	F-t	35.3±1.6	Volcanic breccia	Van Horn (1981)
13	Biotite	K-Ar	38.8±1.2	Vocanic breccia	Van Horn and Crittenden (1987)
14	Biotite	K-Ar	38.3±1.1	Traverse Volcanics of Slenz (1955)	Crittenden and others (1973)
15	Whole rock	K-Ar	21.4±2.5	Basalt	Moore and McKee (1983)
16	Biotite	K-Ar	34.8±1.3	Keetley Volcanics	Bromfield and others (1977)
16	Hornblende	K-Ar	37.5±1.3	Keetley Volcanics	Bromfield and others (1977)
17	Biotite	K-Ar	36.0±1.1	Keetley Volcanics	Crittenden and others (1973)
18	Biotite	K-Ar	33.6±1.0	Keetley Volcanics	Crittenden and others (1973)
19	Biotite	K-Ar	34.9±1.0	Keetley Volcanics	Crittenden and others (1973)
20	Phlogopite	K-Ar	37.6	Keetley Volcanics	Best and others (1968)
20	Phlogopite	K-Ar	38.8	Keetley Volcanics	Best and others (1968)
21	Biotite	K-Ar	38.5	Norwood Tuff	Evernden and others (1964)
21	Sanidine	K-Ar	38.4	Norwood Tuff	Evernden and others (1964)

caldera (Morris, 1975; Morris and Lovering, 1979). North of Utah Lake, the Traverse Mountains are partly formed by latitic volcanic rocks (Moore, 1973).

Small areas of tuff and volcanic breccia of middle Tertiary age are exposed just north of Salt Lake City in the Salt Lake salient, a bedrock protrusion into the Salt Lake Valley that is bounded by Neogene faults on all sides (Van Horn, 1981; Van Horn and Crittenden, 1987). Intermediate volcanic rocks and lahars and a few small intrusions are exposed in the Stanbury Mountains (Moore and McKee, 1983).

RADIOMETRIC AGES OF MIDDLE TERTIARY VOLCANIC ROCKS

Relatively few isotopic ages have been determined on the Keetley Volcanics. A few ages have been determined by using the K-Ar method; these are listed in table 1, and sample localities are shown on plate 1. The ages range from 33.6 Ma for biotite at locality 10 to 34.8±1.3 for biotite and 37.5±1.3 Ma for hornblende at locality 16. The Keetley Volcanics are cut by the Indian Hollow plug and the Park Premier stock, which gave hornblende K-Ar ages of 37.0±1.3 and 36.1±1.0 Ma, respectively. A concordant age of 34.8±1.2 Ma for biotite from the Park Premier stock (Bromfield and others, 1977) tends to confirm the hornblende ages.

The ages of the stocks in the Wasatch Range west of the Keetley Volcanics are less well defined because of discordance between the biotite and hornblende ages. Bromfield and others (1977) considered these hornblende ages to be too old because of excess argon or xenocrystic hornblende. The biotite ages for the stocks just east of the area of the Keetley Volcanics (shown as one unit on plate 1) are similar to those of the Keetley Volcanics (fig. 2).

The stocks in the western part of the Wasatch Range are apparently younger than those in the eastern part. K-Ar ages of biotite and fission-track ages of zircon and sphene obtained for the Alta stock are concordant at about 33 Ma (Crittenden and others, 1973), whereas K-Ar ages of hornblende range from 31.7±1.5 to 47.8±1.9 Ma (Crittenden and others, 1973; Bromfield and others, 1977). Small intrusions (not shown on plate 1) northwest of the Alta stock and north of the Little Cottonwood batholith have biotite K-Ar ages of from 32.5±0.9 to 36.7±0.9 Ma (James and McKee, 1985). The westernmost intrusion in the Wasatch Range, the Little Cottonwood batholith, has zircon and sphene fission-track ages of from 23.1±3.0 to 25.1±3.2 Ma, biotite K-Ar ages of from 25.1±6 to 29.1±1.4 Ma, a muscovite K-Ar age of 26.2±0.8 Ma, and a hornblende K-Ar age of 31.9±0.9 Ma. Parry and Bruhn's (1986) study of the southwestern corner of the Little Cottonwood batholith indicates that the batholith was altered during an early stage of movement on the Wasatch fault zone when it was apparently at a depth of 10 km and a temperature of more than 300 °C. Hydrothermal sericite from the zone of altered and deformed rock in the fault zone has a K-Ar age of 17.6±0.7 Ma (Parry and Bruhn, 1986). In view of this information, all or some of the isotopic ages could be cooling ages related to uplift of the Wasatch Range. Some of the samples

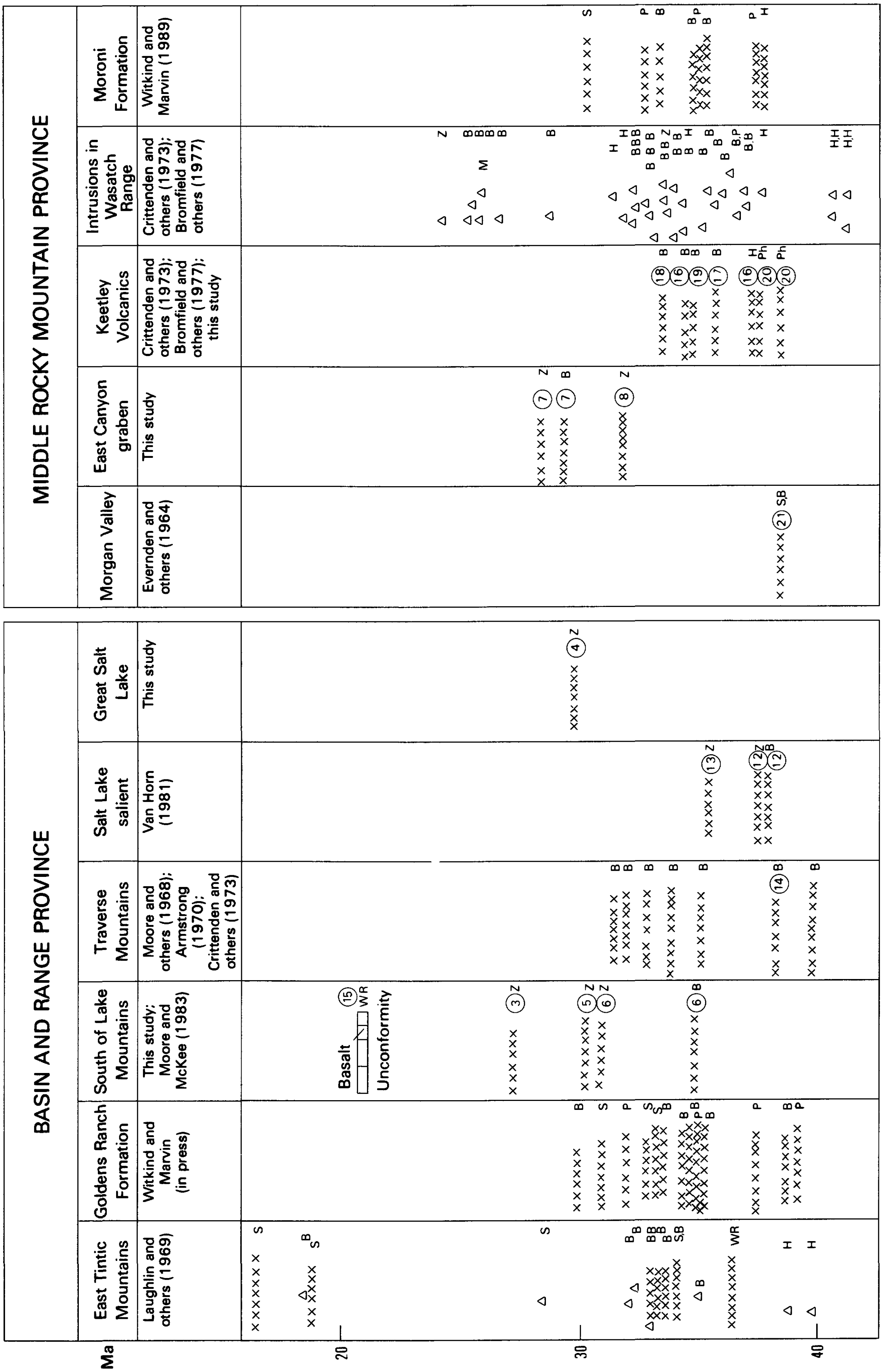

Figure 2. Correlation of middle Tertiary rocks based on isotopic ages. ×, extrusive rock; Δ, intrusive rock. Zircons dated by fission-track method; all other minerals by K-Ar method. Minerals: B, biotite; H, hornblende; M, muscovite; P, plagioclase; Ph, phlogopite; S, sanidine; WR, whole rock; Z, zircon. Number in circle corresponds to locality number of dated sample shown on plate 1.

Table 2. Fission-track ages of zircon and apatite from tuffs in the western part of the Salt Lake City 1° × 2° quadrangle and vicinity

[Sample localities shown on plate 1. Constant: $\lambda_f = 7.03 \times 10^{-17} yr^{-1}$. Number of tracks (t) counted is given in parentheses. Analyst: C.W. Naeser]

			Density			Age	Number	Uranium	Location	
Loc.	Field number	Laboratory number	Fossil tracks $\times 10^6$(t/cm^2)	Induced tracks $\times 10^6$(t/cm^2)	Neutron dose $\times 10^{15}$(n/cm^2)	$\pm 2\sigma$ ((Ma)	of grains counted	content (ppm)	Latitude (north)	Longitude (west)
				Basin and Range province						
1	Harper Pit	DF-3954	0.810 (105)	10.97 (711)	0.993	4.4±1.0	6	350	40°39'30'	112°01'45'
2	AI-81-1	DF-3952	.54 (159)	5.14 (755)	.973	6.1±1.1	14	170	40°54'44"	112°10'31"
3	SP-10	DF-4958	4.09 (1022)	6.78 (847)	.756	27.2±2.8	6	280	40°11'17"	111°59'00"
4	Am-3	DF-3225	3.62 (537)	7.47 (553)	1.03	29.9±3.9	6	230	41°09'42"	112°33'30"
5	SP-6	DF-2895	5.38 (1022)	10.98 (1042)	1.03	30.2±1.5	6	310	40°12'14"	111°58'39"
6	SP-14	DF-2896	6.48 (1080)	13.00 (1083)	1.03	30.7±1.3	6	360	40°10'26"	111°57'23"
		DF-2896[1]	.007 (145)	.094 (196)	1.03	45.5±12.0	50	2.6		
				East Canyon graben and vicinity						
7	E-31	DF-4435	3.81 (740)	7.45 (724)	.936	28.6±3.2	6	250	40°54'56"	111°34'52"
8	BD-30-b	DF-4788	4.74 (1141)	7.56 (910)	.856	32.0±3.2	6	280	40°51'07"	111°37'19"
9	BD-53	DF-4960	3.06 (652)	4.14 (441)	.756	≥33.4±4.4	6	170	40°49'06"	111°31'03"
10	BD-52	DF-4959	5.20 (1132)	6.53 (710)	.756	≥36.0±3.9	7	270	40°49'56"	111°30'12"

[1]Apatite.

dated by Crittenden and others (1973) are from the eastern margin of the Little Cottonwood batholith; those samples must have crystallized at a significantly shallower depth than those from near the Wasatch fault. The Wasatch Mountains have been tilted eastward and are overlain by Keetley Volcanics along their eastern margin. Thus, the K-Ar ages of the Alta stock and Little Cottonwood batholith may be very similar to their emplacement ages. Although the precise age of the Little Cottonwood batholith is not known, the batholith is probably middle or late Oligocene and younger than the early Oligocene stocks farther east.

South of the Spanish Fork River, K-Ar ages of the Moroni Formation within the area shown in plate 1 range from 32.8±3.3 Ma for plagioclase to 37.8±2.2 Ma for hornblende. The total range of K-Ar ages of minerals from the Moroni is from 30.6 to 37.8 Ma, disregarding a spurious age of 74.6 Ma given by a plagioclase (Witkind and Marvin, 1989). The correlative Goldens Ranch Formation, south of the Oquirrh Mountains, has K-Ar ages of from 29.9±1.1 to 38.8±0.9 Ma (fig. 2).

K-Ar ages for volcanic and intrusive rocks in the Tintic Mountains range from 28.5±0.8 to 39.7±1.9 Ma (Laughlin and others, 1969; Morris and Lovering, 1979). The maximum age, for hornblende, is paired with a biotite age of 32.9±0.9 Ma; apparently the hornblende age is too old because the dated rock cuts volcanic rocks that have concordant biotite and sanidine ages of 33.6±1.0 and 33.5±1.0 Ma, respectively (Laughlin and others, 1969; Morris and Lovering, 1979). Age data (Laughlin and others, 1969; Morris and Lovering, 1979) indicate that the East Tintic igneous center was active 32–34 Ma, and that minor igneous activity occurred again about 18.5 Ma (fig. 2).

In the Oquirrh Mountains, ages determined for minerals from fresh and hydrothermally altered intrusive rocks in the Bingham mining district all fall within a narrow time span of 36.6±0.3–39.8±0.4 Ma (Moore and others, 1968; Moore and Lanphere, 1971; Warnaars and others, 1978). Ages from the volcanic field in the western Traverse Mountains, however, cover a wider range. Some of the flows in that field are intruded by a rhyolite plug, which has a biotite age of about 34 Ma (Moore and others, 1968). The youngest flows in the volcanic pile are about 32 Ma (Moore, 1973), and the oldest has biotite ages as old as 39.8±0.9 Ma (Armstrong, 1970). In the eastern Traverse Mountains, a biotite sample from the volcanic rocks has an age of 38.3±1.1 Ma (loc. 14) (Crittenden and others, 1973).

Biotite ages from a volcanic breccia in the Salt Lake salient (loc. 12) are 37.7±1.1 and 38.8±1.2 Ma (loc. 13); zircon from the breccia at locality 12 has a fission-track age of zircon 35.3±1.6 Ma. An underlying tuff near locality 12 yielded zircon that has a fission-track age of 37.4±1.6 Ma and biotite that has a K-Ar age of 45.0±1.4 Ma. The latter age is apparently too old (Van Horn, 1981).

In this study, we dated several zircons and one biotite from volcanic rocks in the East Canyon graben and at the base of the Keetley Volcanics to the south (tables 2, 3). A tuff at or near the base of the conglomerate (loc. 10) that interfingers with the Keetley Volcanics has a minimum zircon fission-track age of 36.0±3.9 Ma. This age is similar to the ages from the main body of the Keetley Volcanics (table 1). A tuff at about the same stratigraphic horizon (loc. 9) has a

Table 3. K-Ar ages of biotite from tuffs in the western part of the Salt Lake City 1° × 2° quadrangle

[Sample localities shown on plate 1. Potassium analyses made using a lithium internal standard. Constants: $^{40}K\lambda_{\varepsilon}=0.581\times10^{-10}yr^{-1}$, $\lambda_{\beta}=4.962\times10^{-10}yr^{-1}$. Atomic abundances for $^{40}K/K$ total=1.167×10^{4} mol/mol. Analysts: R.F. Marvin, H.H Mehnert, V.M. Merritt, and E.L. Brandt]

Locality	Field no.	K_2O (wt. percent)	Radiogenic ^{40}Ar (moles/g[1])	Radiogenic ^{40}Ar (percent[2])	Age ±2σ (Ma)
6	SP-14	7.43	3.770	87	34.9±1.2
			3.739	75	34.6±1.2
7	E-31	4.815	2.067	70	29.6±.1

[1]Times 10^{-10}.

[2]Of total argon.

minimum fission-track age of 33.4±4.4 Ma, the error limits of which overlap with the age for locality 10.

In the East Canyon graben (loc. 8), a hard fresh part of an ash-flow tuff, 200 m or less above the base of the volcanic sequence, has a zircon fission-track age of 32.0±3.2 Ma. Biotite and zircon from a crystal-lithic tuff about 600 m above the base of the volcanic deposits (loc. 7) have ages of 29.6±1.1 and 28.6±3.2 Ma, respectively. Although the ages are concordant, the biotite has a low K_2O content that makes its calculated age somewhat suspect. Samples from both localities 7 and 8 are younger than the few available ages from the main body of the Keetley Volcanics but are in the range of age for the Alta stock and the Little Cottonwood batholith. Those intrusions may be the roots of volcanoes that produced the rocks dated at localities 7 and 8 (fig. 2).

Only one sample of Norwood Tuff in the Morgan Valley (loc. 21) has been dated. The sample was collected about 800 m above the base of the Norwood, and its age of 38.5 Ma is based on both biotite and sanidine ages (Everden and others, 1964). Stratigraphic thickness calculations, based on limited structural data in the Morgan Valley (Mullens and Laraway, 1973; R.J. Hite, U.S. Geological Survey, written commun., 1977; Bryant, 1984, unpub. data), suggest that the Norwood Tuff may be as thick as 4.5 km; however, undetected folds and faults may be present so that the true thickness of the Norwood Tuff could be less. The apparent substantial thickness of the Norwood Tuff and the ages of related rocks in the East Canyon graben suggest that the upper part of the Norwood is significantly younger than 38.5 Ma.

The great apparent thicknesses of both the distal Norwood Tuff in the Morgan Valley and the transitional-facies rocks in the East Canyon graben suggest that some downwarping or faulting occurred during deposition. The orientation and size of the structures controlling deposition are not known because the present distribution of the basin-filling rocks is controlled by Neogene faulting and folding. The distribution and ages of rocks, as understood in the 1950's, led Eardley (1955) to suggest that mid-Tertiary folding or faulting along northerly trends may have occurred locally. Detailed stratigraphic and sedimentological studies of these deposits might produce a more definitive answer to the cause of their large apparent thicknesses. Such studies, however, will be handicapped by poor exposures, especially near the major faults bordering the valleys where the thickest sections are inferred from the dips of the beds.

The south end of the Lake Mountains, known as the Fox Hills, is a structurally and topographically low area that connects the more rugged, main part of the Lake Mountains with the East Tintic Mountains to the south. In this area, tuff, claystone, and limestone are capped by potassic basalt, and the base of the section is nowhere exposed. The stratigraphically lowest unit exposed in most of the area is a crystal-vitric tuff rich in pumice fragments. At one location, a welded tuff crops out below the pumice-rich tuff. Lithologically, these rocks resemble those in Rush Valley to the northwest, which were described by Heylmun (1965) and assumed to be of late Miocene or Pliocene age. These rocks were formerly assigned to the Salt Lake Formation (Bullock, 1951). Similar rocks exposed in the Jordan River Canyon between the eastern and western Traverse Mountains form part of the Jordan Narrows unit of the Salt Lake Group of Slentz (1955). Samples of the tuff beds from localities 5 and 6 have zircon fission-track ages of 30.2±1.5 Ma for the pumice and 30.7±1.3 Ma for the welded tuff (table 2). Biotite from the welded tuff has a K-Ar age of 34.75±1.2 Ma (average of two almost identical values, table 3). The correlation of this tuff horizon with possible sources in the region is uncertain. If the biotite age is correct, the tuff may correlate with the Packard Quartz Latite of the East Tintic Mountains, which has K-Ar ages of 33.6±1.0 (biotite) and 33.5±1.0 Ma (sanidine) (Laughlin and others, 1969). The Packard Quartz Latite may have erupted during a caldera-forming event, and airfall tuffs related to it may be widely distributed (Morris, 1975; Morris and Lovering, 1979). If the zircon ages are correct, correlation is less certain. None of the major plutons of the East Tintic Mountains have been dated, but a small latite porphyry intrusion has given a K-Ar sanidine age of 28.5±0.8 Ma (Laughlin and others, 1969). In the western Traverse Mountains, a biotite age of 31.4 Ma was obtained from a latite tuff breccia (Moore, 1973). Thus, it is possible that the lower tuffs in the Fox Hills correlate with a volcanic unit younger than most of the extrusive rocks in the East Tintic Mountains.

Approximately 100 m above the tuffs dated in the Fox Hills section (loc. 3), a clay bed as thick as 6 m, informally called the Fox clay (Stringham and Sharp, 1950), is believed to have formed by hydrothermal alteration of a tuff (Ames and Sand, 1957). Available evidence from geologic mapping suggests that this deposit has never been deeply buried, so we think it unlikely that the alteration necessary to form clay from the tuff occurred at high enough temperatures to anneal fission tracks in zircon. Euhedral zircons from the clay have a fission-track age of 27.2±2.8 Ma. This age indicates that the tuffs at the south end of the Lake Mountains are a part of the middle Tertiary volcanic sequence and probably a few million years younger than those of the East Tintic Mountains. We correlate the tuffs at the south end of the Lake Mountains with the Goldens Ranch Formation rather than the Salt Lake Formation. The uppermost unit exposed at the south end of the Lake Mountains is a potassic basalt (Bullock, 1951) that has a K-Ar whole-rock age of 21.4±2.5 Ma (loc. 15) (Moore and McKee, 1983).

Through the kindness of Amoco Production Company, we obtained core samples from a deep exploration well (Bridge State of Utah) in the Great Salt Lake. In this well (loc. 4), an unconformity occurs at a depth of 3,130 m between an upper sequence of greenish-gray and gray tuffs and sediments, previously dated by Naeser and others (1983) as 10.3 million years old at a depth of 2,721 m (table 1), and a lower sequence of reddish-brown and brown tuffs and sediments. Zircons from a tuff at a depth of 3,674 m (4 m above the bottom of the hole) have a fission-track age of 29.9±1.3 Ma, an age comparable to ages of middle Tertiary rocks at the south end of the Lake Mountains and in the East Canyon graben (fig. 2). The unconformity between these two suites of Tertiary rocks apparently was not detected in the seismic study of the Great Salt Lake area (Bortz and others, 1985); therefore, the distribution of the older, mid-Tertiary volcanic rocks and sedimentary rocks in the deep grabens beneath the Great Salt Lake is not known. Other wells in the area that penetrated pre-Tertiary rocks apparently did not encounter middle Tertiary volcanic rocks (Bortz and others, 1985).

A seismic refraction line adjacent to and east of the southeast corner of Great Salt Lake reveals two different velocity layers above a layer presumed to represent consolidated pre-Tertiary rock (Arnow and Mattock, 1968). The middle layer was interpreted by Arnow and Mattock (1968) as Tertiary and the upper layer as Quaternary; however, "about 300 feet of volcanic rock, possibly andesite" was found in a well that penetrated the middle layer (Arnow and Mattock, 1968). Most volcanic rocks in this area, except for light-colored tuffs, are of middle Tertiary age, and the middle Tertiary rocks probably are present in the well.

The original areal distribution of middle Tertiary volcanic rocks apparently was not affected by the basins and ranges existing today. Volcanic materials must have been extruded from numerous centers in the Wasatch, Oquirrh, Tintic, and Stansbury Mountains (Moore and McKee, 1983), and possibly from centers now buried beneath the Neogene basin fill. Volcanic flows and tuffs and associated sediments filled preexisting valleys and spread over the existing topography (Morris and Lovering, 1979). Major drainages leading north and southeast from the Keetley volcanic center carried volcanic detritus many kilometers from the volcanic edifices. Detritus from volcanoes in the East Tintic Mountains was carried southward to form the Goldens Ranch Formation and eastward and southeastward to form the Moroni Formation. Later, during formation of Basin and Range topography, a process continuing to the present, much of the middle Tertiary volcanic rock was eroded from the strongly uplifted blocks that form the Wasatch, Oquirrh, and Stansbury Mountains. Volcanic rocks are preserved in less uplifted blocks, such as the Salt Lake salient, the Traverse Mountains, the tilted flank of the Stansbury Mountains, and the East Tintic Mountains. Volcanic rocks are especially thick in the East Tintic Mountains where they fill a caldera (Morris, 1975; Morris and Lovering, 1979). They are also preserved in Neogene structural lows, such as the saddle between the Wasatch and Uinta Mountains, and in deep basins such as Great Salt Lake. In the area of the Great Salt Lake, north-flowing drainages may have carried volcanic material away from the centers in the Oquirrh and Stansbury Mountains.

A change from predominantly calc-alkalic igneous activity to predominantly basaltic igneous activity may be indicated by the potassic basalt, dated as 21 million years old (loc. 15; table 3), at the top of the section at the south end of the Lake Mountains. Does extrusion of this basalt indicate the onset of crustal extension in the eastern Basin and Range, as suggested by Christiansen and Lipman (1972)? The basalt at the south end of the Lake Mountains is a potassium-rich mafic lava, as defined by Best and others (1980). In southwest Utah, this type of lava was extruded in early Miocene time and was followed after a long time gap by extrusion of tholeiitic basalt in late Miocene time (Best and others, 1980). Moore and McKee (1983) placed the beginning of Basin and Range faulting at 12–13 Ma near the Stansbury Mountains because of nearly concordant ages of petrographically similar olivine basalts exposed on the crest and at the margin of that range. The K-Ar age of hydrothermal sericite along the Wasatch fault at the southwestern corner of the Little Cottonwood batholith suggests that movement on that fault was already in progress 17 Ma (Parry and Bruhn, 1986). Extrusion of potassium-rich mafic lavas may mark the onset of regional extension considerably earlier than the Basin and Range faulting that probably is a reflection of more rapid extension since the late Miocene.

NEOGENE DEPOSITS

Neogene deposits crop out at isolated localities along the margins of the basins occupied by Great Salt Lake and Utah Lake and have been penetrated in drill holes within the basins. Many of the deposits at the basin margins, however, are coarse conglomerates and thus difficult or impossible to date.

In the Salt Lake salient, conglomerate and sandstone, interpreted to be 2.6 km thick, unconformably overlie middle Tertiary volcanic rocks in City Creek Canyon (Van Horn and Crittenden, 1987). Paleocurrent studies by Mann (1974) show that these rocks, which Mann correlated with the Wasatch Formation, were derived from the east. The deposits contain clasts derived both from the Wasatch Formation that caps the crest of the Wasatch Range to the east and from Paleozoic and Precambrian rocks exposed nearby. The position of this conglomerate along the margin of the range, its content of locally derived clasts, and the paleocurrent data suggest that it was a thick range-front deposit that was uplifted, locally tilted, and stranded on an intermediate-level structural block, possibly during a change in the pattern of active faulting along the west front of the Wasatch Range. No material suitable for dating by isotopic or other methods has yet been found in this deposit.

Slentz (1955) correlated several units exposed along the Jordan River and a tributary in the Jordan Narrows area with the Salt Lake Formation. They are, from youngest to oldest: (1) the Harkers Fanglomerate, consisting of conglomerate and sandstone in a tuffaceous matrix, siltstone and claystone, and a unit of travertine (believed by Slenz to be contemporaneous with the Harkers), in which the remains of a fossil horse of Blancan (Pliocene) age were found; (2) the Camp Williams unit, composed of red to tan mudstone, siltstone, impure sandstone, and a basal conglomerate; and (3) the Jordan Narrows unit, composed of white marlstone, limestone, sandstone, and tuff. According to Slentz (1955), structural discordances and erosional unconformities separate the various units. A middle Miocene to middle Oligocene age was assigned to the Jordan Narrows unit by Slentz (1955) because the unit was thought to interfinger with middle Tertiary volcanic rocks of the Traverse Mountains. Ostracods of Miocene to Pliocene age in the Jordan Narrows unit (Slenz, 1955) and a zircon fission-track age of 6.5 Ma from a crystal-vitric tuff in the same unit (loc. 11), however, indicate that all the deposits assigned to the Salt Lake Formation in the Jordan Narrows area are late Miocene or younger in age.

The Harkers Fanglomerate as used by Slentz (1955) was renamed the Harkers Alluvium by Tooker and Roberts (1971) and given a Pleistocene age because of its unconsolidated nature and its position above known Tertiary deposits. Tooker and Roberts (1971) mapped it as a discontinuous apron around the Oquirrh Mountains where it consists of deeply dissected fans exposed topographically above the upper Pleistocene deposits of Lake Bonneville. They characterized the Harkers Alluvium as poorly sorted boulder conglomerate, coarse to fine gravel, silt, and clay. W.E. Scott (written commun., U.S. Geological Survey, 1980) sampled a lens of tuff in a fan gravel, probably in the proximal facies of the Salt Lake Formation, exposed in the Harper gravel pit (loc. 1) about 4 km northeast of the type locality of the Harkers in Harkers Canyon on the east flank of the Oquirrh Mountains (Slentz, 1955; Tooker and Roberts, 1971) and only 1 km from the Harkers mapped in a small exposure (Tooker and Roberts, 1971). Zircon from this tuff gave a fission-track age of 4.4±1.0 Ma (loc. 1, table 2). Lithologically, this deposit resembles the Harkers Alluvium at its type locality, and it seems reasonable to correlate rocks at the gravel pit with those at the type locality. However suggestive, the dated rocks may not be correlative with Harkers of the type locality.

North of Alpine, at the east end of the eastern Traverse Mountains, conglomerate and debris flows containing volcanic and quartzite clasts overlie middle Tertiary volcanic rocks. Much of the material was derived from quartzite and calcareous quartzite of the Pennsylvanian and Permian Oquirrh Group, which probably once overlaid the southern part of the Little Cottonwood batholith. High topographic relief may have been caused by the intrusion of the batholith or by faulting in early Miocene time. Parry and Bruhn (1986) indicated that about 10 km of Neogene uplift occurred on the Wasatch fault just north of the eastern Traverse Mountains and that uplift seems the most likely cause of steep topography from which debris flows and conglomerate could have been derived. No datable material has yet been found in these deposits. Bullock (1958) mapped part of this deposit and assigned a Pleistocene age to it. It lacks much debris from the adjacent Little Cottonwood batholith, which forms the steep mountain front immediately to the east, and therefore it seems reasonable that it is considerably older than Pleistocene.

Other outcrops of coarse deposits of probable Neogene age occur in the area. There are some deeply dissected angular conglomerates of probable Pliocene or early Quaternary age (M.N. Machette, U.S. Geological Survey, oral commun., 1987) south of the Provo River along the west front of the Wasatch Range. Along the mountain front south of Utah Lake, conglomerate, sandstone, and clayey limestone exposed in a fault block dip as much as 20° S. into the main strand of the Wasatch fault zone (M.N. Machette, oral commun., 1987). At Little Mountain at the mouth of Payson Canyon, a cobble-and-boulder conglomerate contains clasts that were derived from the Wasatch block during Neogene or early Quaternary time after uplift of the Wasatch Range. All these deposits may be proximal facies of the Salt Lake Formation.

White tuff exposed below late Pleistocene deposits in a gravel pit on the east flank of Antelope Island (loc. 2) is associated with conglomerate, sandstone, and claystone. Zircon from the tuff has a fission-track age of 6.1±1.1 Ma. In the Amoco Production Company Bridge State of Utah well (loc. 4) in the central graben of the Salt Lake Valley, a crystal-vitric tuff from a depth of 2,721 m has a zircon fission-track age of 10.3±1.0 Ma. If we project a constant rate of filling of the graben (based on the age and depth of the vitric tuff sample) down to the unconformity above the middle Tertiary rocks at 3,130 m depth, we obtain an age of 11.8±1.0 Ma for the oldest basin-fill material deposited in the graben. This age would be a minimum for inception of Basin and Range faulting, and such projections in this tectonic environment are semiquantitative at best.

In the Morgan Valley, a unit of conglomerate as thick as 300 m that contains lenses of tuffaceous limestone dips gently and unconformably overlies moderately dipping Norwood Tuff (Mullens and Laraway, 1973). The lithology of this conglomerate and its structural relations with the middle Tertiary Norwood Tuff suggest that it is correlative with the Salt Lake Formation of the Basin and Range province.

CONCLUSIONS

Available data indicate that the topography of the Salt Lake City region from late Eocene through Oligocene time differed significantly from that during late Miocene time to the present. Flows, tuffs, and conglomerates emanating from numerous volcanic centers both east and west of the Wasatch front formed aprons or filled valleys. Prevailing westerly winds carried tuffs eastward into the Uinta basin (Bryant and others, this volume). The youngest extrusive mid-Tertiary rocks are about 27 Ma.

By about 10 to 12 Ma, the gross pattern of present-day Basin and Range topography was established.

Potassic basalt extruded some 21 million years ago may signal the change in structural pattern and the onset of regional extension, and movements on some Basin and Range faults occurred shortly thereafter. The age of the oldest dated deposit confined to a Neogene basin is 10 Ma, an age compatible with fission-track ages determined on apatite in the Wasatch Range that indicate rapid uplift of the Wasatch block during the past 10 million years (Naeser and others, 1983). A more precise determination as to when rapid extension began in this area remains elusive.

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Appendix. Description of sample localities

1. Lens of white tuff in alluvial-fan gravel beneath Lake Bonneville deposits exposed in Harper gravel pit. Proximal facies of the Salt Lake Formation, possibly correlative with the Harkers Alluvium. Collected by W.E. Scott. Below power line at 4,780-ft contour. SE1/4 sec. 11, T. 2 S., R. 2 W., Salt Lake County. Magna quadrangle.
2. Very light greenish gray vitric tuff. Sparse crystal fragments of quartz and plagioclase in a matrix of glass shards. Collected by Bruce Bryant. From floor of Utah State Highway Department gravel pit 1,050 m S. 53° E. of Mushroom Spring on Antelope Island. NW1/4 sec. 15, T. 2 N., R. 3 W., Salt Lake County. Antelope Island quadrangle.
3. Red clay bed, 0.8 m thick, overlying light-yellowish-brown-weathering white clay and overlain by limestone. Fox clay deposit (Stringham and Sharp, 1950; Ames and Sand, 1957). The interpretation that the deposit is an altered tuff (Ames and Sand, 1957) is supported by the euhedral shape of the zircons obtained from it. Collected by Bruce Bryant. From claypit 560 m S. 60° W. of 5,340-ft-high hill south of center of sec. 20, T. 7 S., R. 1 W., Provo County. Soldier Pass quadrangle.
4. Andesitic tuff at depth of 3,675 m in a core of Amoco Production Company Bridge State of Utah Well in the Great Salt Lake. Angular crystal fragments of andesine and a few of biotite and quartz in a matrix of cryptocrystalline material and glass. Contains a few carbonate aggregates (possibly rock fragments) as long as 3 mm. Collected by Amoco Production Company. NW1/4 sec. 12, T. 5 N., R. 7 W., Davis County. Carrington Island NE quadrangle.
5. Crystal vitric tuff from near base of exposed section at south end of Lake Mountains. Fragments of andesine and euhedral biotite crystals in a matrix of pumice fragments and glass. Contains sparse lithic fragments. Collected by Bruce Bryant. At 5,100 ft altitude, north side of road west of Soldiers Pass. Near center of south edge of sec. 17, T. 7 S., R. 1 W., at "pumice mine," Utah County. Soldiers Pass quadrangle.
6. Welded tuff composed of crystals and fragments of plagioclase and biotite and fragments of squashed pumice in a glassy matrix. From near base of exposed section at south end of Lake Mountains. Collected by Bruce Bryant. At 4,830 ft altitude, low ridge south of road. Sec. 28, T. 7 S., R 1 W., Utah County. Soldiers Pass quadrangle.
7. Crystal lithic tuff. Fragments of andesite in a very fine grained matrix containing fragments of plagioclase, brown hornblende, and biotite. Collected by Bruce Bryant. Roadcut on Utah Highway 65, 1,970 m N. 87° E. from Saddle Rock. Sec. 11, T. 2 N., R. 3 E., Morgan County. East Canyon Reservoir quadrangle.
8. Andesitic ash-flow tuff. Crystals and fragments of andesine, green hornblende, biotite, and monoclinic pyroxene in a matrix of cryptocrystalline material and glass. Collected by Bruce Bryant. South side of Utah Highway 65, in Little Dutch Hollow at 6,520 ft altitude. Sec. 4, T. 1 N., R. 3 E., Morgan County. Little Dutch Hollow quadrangle.
9. Tuff composed of fragments of quartz, plagioclase, biotite, monoclinic pyroxene, and green hornblende and fragments of sedimentary rock in a matrix of altered glass. Near base of conglomerate that is below and interfingers with Keetley Volcanics and unconformably overlies Wasatch Formation. Collected by Bruce Bryant. From roadcut, 200 m S. 50° W. of corner in Morgan-Summit County line on 8,400-ft knob southeast of center of sec. 17, T. 1 N., R. 4 E., Morgan County. Big Dutch Hollow quadrangle.
10. Tuff composed of fragments of quartz, plagioclase, hornblende, biotite, pumice, and sedimentary rock in a matrix of clay and carbonate. Near base of conglomerate that is below and interfingers with Keetley Volcanics and unconformably overlies Wasatch Formation. Collected by Bruce Bryant. On gentle ridge, 360 m S. 88° W. of Morgan-Summit County line and east border of Big Dutch Hollow quadrangle. Sec. 9, T. 1 N., R. 4 E., Morgan County.

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