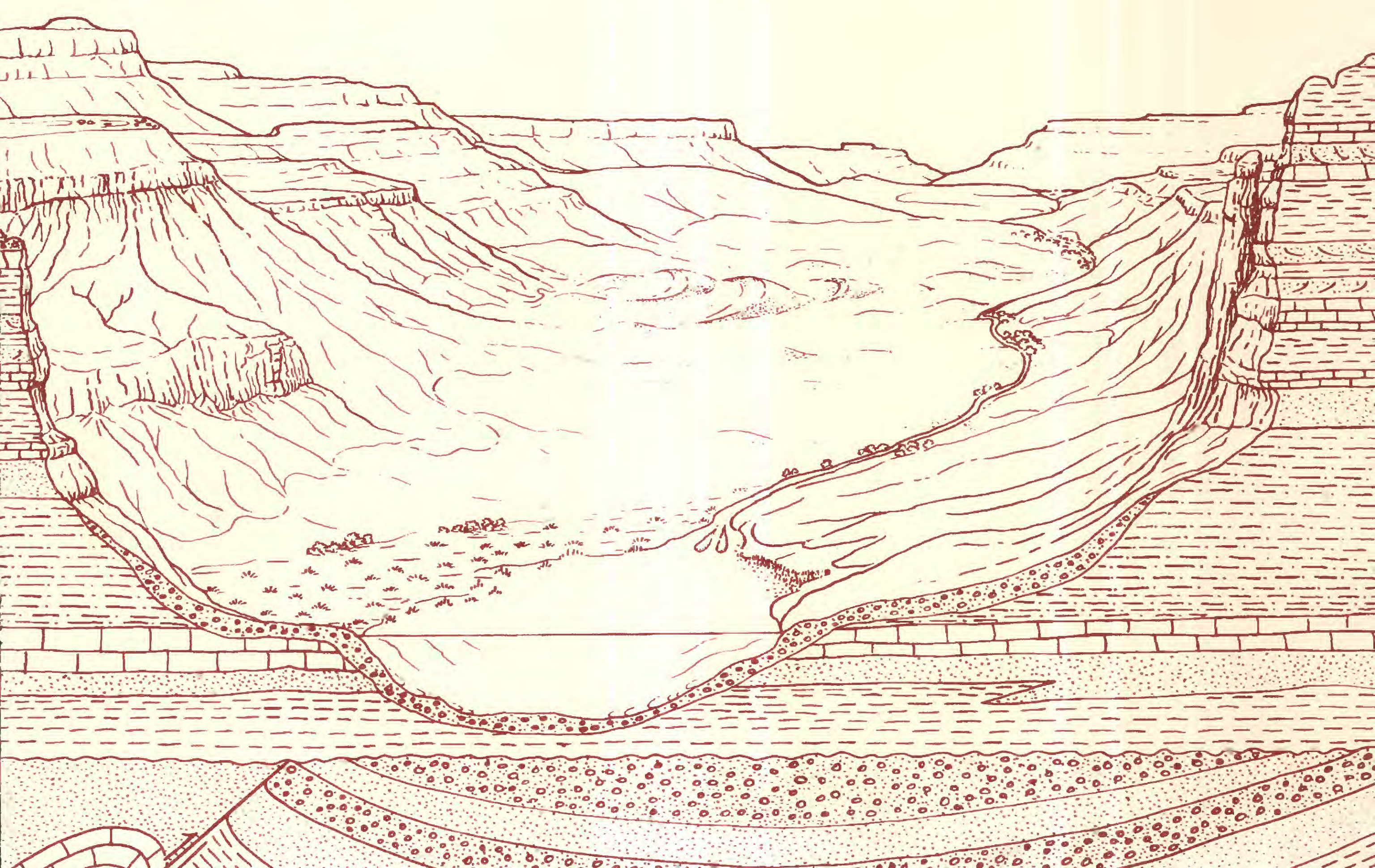
Geometry and Structural Evolution of Gilsonite Dikes in the Eastern Uinta Basin, Utah

Prepared in cooperation with the U.S. Department of Energy

U.S. GEOLOGICAL SURVEY BULLETIN 1787–HH



ERRATA

U.S. Geological Survey Bulletin 1787–HH, Geometry and structural evolution of gilsonite dikes in the eastern Uinta basin, Utah.

In the fourth paragraph of the abstract, p. HH1, there is an error in the vertical extent of the dikes. The third sentence should read as follows.

> Amounts of water lost are conjectural, but the large size of the dikes (lengths commonly >10 km, vertical extents 0.3–1 km) provides some measure of migration distances.

Chapter HH

Geometry and Structural Evolution of Gilsonite Dikes in the Eastern Uinta Basin, Utah

By EARL R. VERBEEK and MARILYN A. GROUT

Prepared in cooperation with the U.S. Department of Energy

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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TABLE

Geometry and Structural Evolution of Gilsonite Dikes in the Eastern Uinta Basin, Utah

By Earl R. Verbeek *and* Marilyn A. Grout

Abstract

Numerous long, subparallel dikes in the eastern Uinta basin of Utah are filled with a brittle hydrocarbon most commonly known by its trade name, gilsonite. The dikes strike N. 40°–70° W. and are vertical, or almost so; many can be followed for distances of several kilometers or more, and several have mapped lengths exceeding 15 km. The dikes range in thickness from a fraction of a millimeter to about 5 m and are exposed in strata ranging in age from early Eocene to early Oligocene. Previous studies established the source of the gilsonite as the middle Eocene, bitumen-rich marlstone beds (oil shale) of the upper part of the Green River Formation.

Individual fractures bounding dike walls commonly are described as smooth and relatively featureless but nevertheless are complex in detail and display the full complement of surface structures indicative of extensile failure. Plumose structure, in particular, is common where the wallrocks are fine grained and well cemented. Multiple thin, tapering dikelets that diverge from the main dike at low angles are abundant locally and resulted from the filling of twist-hackle faces wedged open by the intruded gilsonite. Dike-bounding fractures in sandstone typically are large, tens to more than a hundred meters long, but fractures more than 15 m long are uncommon in weakly to moderately indurated mudstone higher in the section.

Dike geometries in three dimensions consist of an interconnected network of longitudinal dike segments, cross segments, and sills whose complexity at any stratigraphic level is strongly dependent on rock type. Longitudinal segments parallel to the overall dike trend are the main and in some places the only component, particularly in sandstone where dilation of large, overlapping fractures resulted in lengthy, continuous dikes with locally stepped walls. Small sills and minor cross dikes at right angles to the main trend, however, are not uncommon. Longitudinal dike segments in mudstone, in contrast, commonly are shorter and thinner and repeatedly split and merge to form complex anastomosing networks connected at intervals by short (1 m or less), thin cross segments. Sills extending outward from dikes are common at all stratigraphic levels and attest to high fluid pressures during expulsion of bitumen from the source beds into the dikes. Estimated maximum emplacement depths range from 700 to 1,300 m for the easternmost dikes to as much as 2,500 m for dikes nearer the center of the basin.

The abundant evidence for forceful rather than passive intrusion suggests that the dikes propagated as hydraulic fractures from overpressured, bitumen-rich source beds in the upper part of the Green River Formation. The presence of limonite and calcite as early deposits on dike walls and of continuous alteration rinds (bleached zones) adjacent to fractures discontinuously occupied by gilsonite shows that the dike fractures were conduits for the expulsion of formation water before the source beds were sufficiently mature to generate significant quantities of liquid bitumen. Amounts of water lost are conjectural, but the large size of the dikes (lengths commonly >10 km, vertical extents 1–3 km) provides some measure of migration distances. Much of the gilsonite, too, likely migrated laterally through the dike-fracture system; gilsonite in the easternmost dikes, in particular, probably was derived from unexposed source beds downdip to the northwest and not from the oil shale directly beneath.

We interpret the gilsonite dikes as early products of a period of regional, post-Laramide, northeast-southwest tectonic extension that affected much of the northeastern Colorado Plateau. Initial failure by hydraulic fracture was prompted both by decrease in the magnitude of regional σ_3 and by high pore-fluid pressure in the gilsonite source beds. Continuing tectonic extension resulted in the formation of a regional set of joints that strike almost parallel to the dikes in most areas but at slight to moderate angles to them in others. Dike and joint walls have much different lengths and geometries despite their similar attitudes in many areas, and abutting relations consistently establish the joints as younger. The joints are members of the most prominent and regionally extensive fracture set to have affected the Tertiary rocks; the set extends beyond the confines of the Uinta basin into the Piceance basin of western Colorado and affected a minimum known area of 30,000 km^2. Minor, west-northwest-trending normal faults in both basins are still-later products of the same general deformation.

INTRODUCTION

The Uinta basin of northeastern and north-central Utah has long been noted for its remarkable vein deposits of several types of solid hydrocarbons. By far the most important of these is gilsonite, a black, lustrous, brittle asphaltite that forms scores of northwest-trending dikes cutting Tertiary sedimentary rocks in the eastern part of the basin. The first recorded discovery of this substance is that of Denton (1866), who described several dikes as much as 1.0 m thick and traced one of them for 8 km across the countryside over a vertical distance of 240–300 m. Though the title of Denton's paper mentions only Colorado [Territory], he gives the locality as "near the junction of White and Green Rivers, and probably in Utah," and his description is a clear reference to gilsonite dikes. The earliest recorded use of the substance (though a notably unsuccessful one) was in 1869 when a blacksmith, mistaking it for coal, used some for fuel and almost burned down his shop as molten gilsonite issued from the forge (Pruitt, 1961). Blake (1885, 1890) provided early scientific descriptions of the new hydrocarbon under the name "uintahite," but usage of that term quickly declined in most circles in favor of the marketing name "gilsonite" (Maguire, 1900). The latter term honors Samuel H. Gilson, prospector and tireless promoter of the material's commercial possibilities, who in 1886 began to explore the region and about 1888 opened the first of its many gilsonite mines (Pruitt, 1961). The early history of gilsonite mining in the eastern Uinta basin is a colorful one, involving wholesale trespass of prospectors on Indian lands, the staking of hundreds of illegal claims, the intentional mislocation of a reservation boundary so as to open some of the land to mining, the passage by Congress in 1903 of an act granting legal status to some of the old claims and providing for the sealed-bid sale of mining rights to much additional land, and the construction of one of the West's most unusual railroads to transport the ore; good accounts are given by Crawford (1957), Kretchman (1957), Remington (1959), Pruitt (1961), Covington (1964), and Chenoweth (1985). Today gilsonite is mined from five large dikes in the eastern part of the dike swarm and from one dike farther west; reserves are considered plentiful and the marketing prospects encouraging.

The source beds for gilsonite and the origin of the fractures that contain it have been long debated. Agreement on the first point seems finally to have been achieved through the papers of Hunt and others (1954) and Hunt (1963), who presented convincing evidence that gilsonite was derived from the middle Eocene, bitumen-rich lacustrine beds of the upper part of the Green River Formation. Few today seem inclined to challenge that view. On matters of structure, however, no consensus has been reached, primarily because so little information on the detailed geometry, intrusion mechanics, or wallrock alteration of the dikes had been available until recently. In a recent report (Verbeek and Grout, 1992), we summarized much new evidence on the structural evolution of the dikes and concluded that they originated as large hydraulic extension fractures within overpressured source beds of the Green River Formation before regional jointing of the Tertiary section took place. Monson and Parnell (1992), in a paper released simultaneously, reached almost identical conclusions through careful study of the petrography and diagenesis of the sandstone host rocks. We here update our earlier work and present our findings in full.

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PROPERTIES AND USES OF GILSONITE

Gilsonite is a black, homogeneous, solid hydrocarbon, lustrous when fresh but dull when weathered, that breaks with a pronounced conchoidal to hackly fracture. The black color and brittleness are evident only in mass; the material when pulverized is chocolate brown and tacky and when handled is difficult to remove completely from the skin except through the use of organic solvents. The discovery in the late 1800's of large dikes of this substance prompted numerous investigations of its physical and chemical properties; among the early reports are those of Hayes (1866), Wurtz (1869, 1889), Blake (1885, 1890), Raymond (1889), Day (1895), and Maguire (1900). Gilsonite has a specific gravity of 1.05–1.10, a hardness (Mohs scale) of 2–2.5, and a melting point of 120°–230°C; it is almost completely soluble in carbon disulfide (Ladoo, 1920; Davis, 1957; Pruitt, 1961). These last two properties characterize it as an asphaltite. Gilsonite contains by weight 85–86 percent C, 8.5–10.0 percent H, 2.2–3.4 percent N, 0.2–0.5 percent S, and about 0.2 percent O (Abraham, 1945; Hunt, 1963). Among known, naturally occurring solid hydrocarbons it is relatively high in volatiles, with an average content of about 88 percent (Cornelius, 1984; Khavari-Khorasani, 1984), and as mined contains almost no mineral impurities. Infrared absorption spectra and mass spectrometry analyses (Hunt, 1963) show that gilsonite is a predominantly aromatic asphaltite.

The low melting point, high solubility in oils, and near-perfect insulator qualities of gilsonite historically have lent it to a wide variety of uses. Gilsonite has been distilled and heat-refined to produce, among other products, a high-octane automotive gasoline, railroad and automotive diesel fuels, LPG (liquefied petroleum gas), raw naphtha, lubricating oils, road oils, and asphalt paving products (Crawford and Pruitt, 1963; Kilborn, 1964). Gilsonite has been used to saturate roofing and other building construction papers, as a base for various paints, varnishes, and anticorrosive coatings, and as a component of battery boxes, phonograph records, insulating and waterproofing jackets for underground pipes, and automotive body sealers (Henderson, 1957; Pruitt, 1961; Jackson, 1985). More recent uses are as a stabilizer in nonsmearing rotogravure inks, an additive for oil-well slurries, a binder for wood fibers, and a high-BTU component of certain explosives (American Gilsonite Company, 1989). Gilsonite can also be heated and calcined to produce a metallurgical-grade carbon coke and in recent years has been used in the manufacture of high-purity carbon electrodes for the nuclear power industry (Jackson, 1985).

GEOLOGIC SETTING

The gilsonite dikes of northeastern Utah are exposed within an area of about 3,600 km^2 in the vicinity of the White and Green Rivers, south of the towns of Vernal and Roosevelt (fig. 1). This area is within the east-central part of the Uinta basin, a sharply asymmetric continental intermontane basin along the northern edge of the Colorado Plateau. The Uinta basin, like the Green River basin to the north, the Piceance basin to the east, and the Paradox and San Juan basins to the south, developed through segmentation in early Cenozoic time of a once continuous Cretaceous foreland depression that extended north to south across the entire North American Continent. The foreland, a broad structural trough within which an eastward-thinning wedge of Upper Cretaceous clastic sediments had accumulated, formed as a crustal downwarp during the Sevier (Albian? to late Campanian) orogeny in response to the emplacement of multiple thrust sheets farther west (Armstrong, 1968). Much of the Wasatch Range and parts of the Wasatch Plateau (fig. 1), which border the Uinta basin on the northwest and southwest, respectively, are underlain by these thrust sheets. Marine and coastal sedimentation in the open Cretaceous sea gradually gave way to lacustrine and fluvial deposition in early Cenozoic time as highlands arose within the foreland; between the highlands, sedimentation increasingly became partitioned within rapidly subsiding, internally drained, intermontane basins (Franczyk and others, 1989). Three of these highlands, the San Rafael Swell and Uncompahgre uplift on the south and the Uinta Mountains on the north (fig. 1), throughout much of Cenozoic time defined the depositional limits of the Uinta basin and still persist as topographic highs today. The history of a fourth, the Douglas Creek arch along the eastern edge of the area containing the gilsonite dikes, was recounted recently by Johnson and Finn (1986). An excellent review of sedimentation within the developing Uinta basin, from its inception in Maastrichtian (latest Cretaceous) time through the late Eocene, was given by Franczyk and others (1989).

Much of the area occupied by the gilsonite dikes is underlain by almost flat-lying strata of Eocene and locally Oligocene age. Local relief typically ranges from 50 to 100 m in a moderately dissected and sparsely vegetated landscape of low buttes, small mesas, and dry washes. Near the eastern margin of the basin, however, the landscape is one of steep cliffs and deep (200–300 m) canyons whose intricately dissected forms were colorfully described by one early observer (Eldridge, 1896a, p. 921) as "presenting in general a scene of much desolation and decay." Here, in an area stratigraphically 600–800 m below much of the basin interior, one can examine the basal parts of several gilsonite dikes where they extend downward into the lateral equivalents of their source beds.

STRATIGRAPHIC SETTING

Principal host rocks for the gilsonite dikes are the lacustrine and fluvial strata of the Green River, Uinta, and lower part of the Duchesne River Formations, of middle Eocene through Oligocene age. Older rocks generally are not exposed in the vicinity of the dikes except near the eastern margin of the basin, where two dikes can be followed into the lower Eocene rocks of the Wasatch Formation. Thicknesses and general lithologies for these units are shown in figure 2.

The Eocene rocks constitute the major part of a sequence, more than 3,600 m thick, of lacustrine and associated synorogenic sediments that filled the Uinta basin in early Cenozoic time (Franczyk and others, 1989). For much of the Eocene Epoch, clastic debris shed from basin-margin uplifts accumulated as fluvial and shoreline sediments around an extensive lake (Lake Uinta) that occupied the basin interior. Facies changes

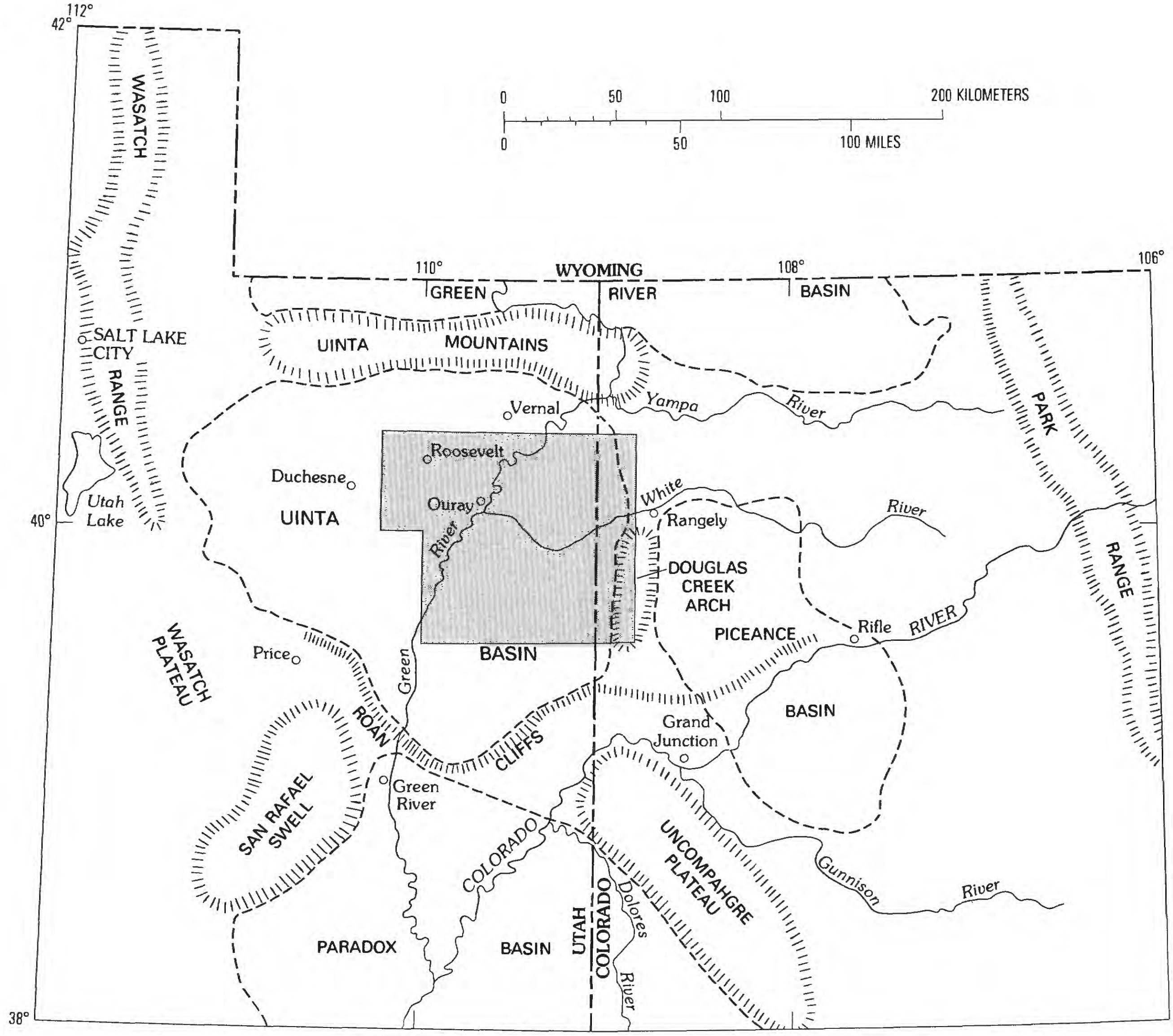

Figure 1. Location of study area (light shading) in east-central part of Uinta basin, Utah and Colorado, in relation to Tertiary basins and uplifts. Modified from Cashion (1967).

from chemically precipitated carbonate rocks to quartzose clastic rocks over distances of only a few kilometers are common (Franczyk and others, 1989), as are complex intertonguing relationships attributed to fluctuating lake levels. The depositional lithofacies that resulted, and from which map units are defined in the field, thus are the partial time equivalents of one another (Hunt, 1963).

Wasatch Formation

Those rocks generally assigned to the Wasatch Formation, of early Eocene age, comprise a predominantly fluvial sequence of red and gray shale and siltstone, fine- to medium-grained gray to brown channel-form sandstone, and local beds of conglomerate deposited by streams draining adjacent highlands (Cashion, 1967; Franczyk and others, 1989). The Wasatch Formation is exposed only in the far eastern part of the study area and lies below the known or suspected bases of most of the gilsonite dikes. In that area, near the basin margin, the formation is fairly thin—Cashion (1967) reported a thickness of only 213 m for one locality—but it thickens appreciably westward, toward the basin interior.

Green River Formation

Fluvial beds of the Wasatch Formation grade laterally, and upward, into marginal-lacustrine rocks of the lower

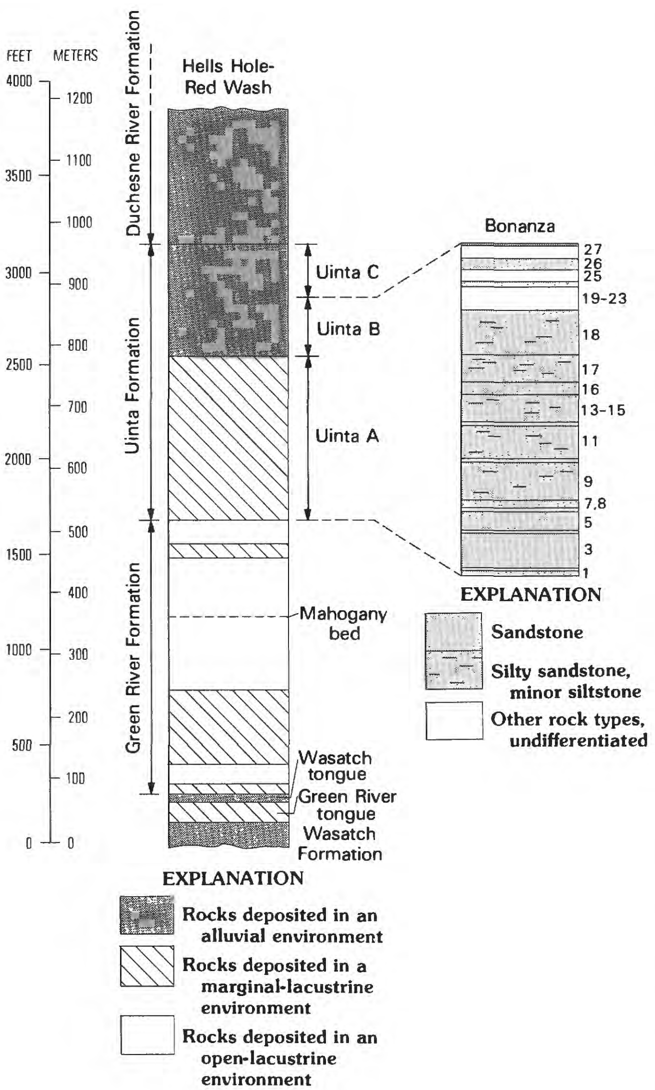

Figure 2. Generalized stratigraphic section (left) of Tertiary rock units in Hells Hole–Red Wash area, modified from section *E–E'* of Franczyk and others (1989). Approximate location of section, about 18 km north of the mining camp of Bonanza, Utah, is shown on plate 1*A*. Partial section at right (same scale) shows thick sandstone units in lower part of Uinta Formation near Bonanza. Numbers to right of section correspond to units measured by Cashion (1982).

part of the Green River Formation, which contains much fine- to medium-grained sandstone, siltstone, and shale along with oolitic, algal, and ostracodal limestone (Cashion, 1967). Abrupt lateral changes in thickness and lithology between the predominantly fluvial beds of the Wasatch Formation and the predominantly shallow water lacustrine Green River beds that intertongue with them reflect the irregular nature of this transition and impede selection of formation boundaries (Cashion, 1967). Subsequent expansion of Lake Uinta in middle Eocene time deepened the lake and extended open-lacustrine conditions over a wide region, resulting in deposition of the fine-grained, bitumen-rich carbonate rocks (oil shale) of the middle and upper parts of the Green River Formation (Bradley, 1931; Johnson, 1985). It is these rocks that generally are acknowledged as the source beds of gilsonite (Hunt and others, 1954; Hunt, 1963; Hatcher and others, 1992). They range in organic content from almost zero to about 50 percent by weight (Cashion, 1957) and contain dolomite, calcite, quartz, potassium feldspar, albite, analcime, and clay minerals as the principal inorganic constituents. Thin interbeds of ash-fall tuff and of limestone, dolostone, calcareous siltstone, and fine-grained sandstone are minor associates (Untermann and Untermann, 1964; Cashion, 1967). The most bitumen-rich part of the oil-shale succession is known as the Mahogany ledge in outcrop (plate 1*A*) and as the Mahogany zone in the subsurface; within this interval is the exceedingly rich Mahogany bed (fig. 2), a prominent regional marker horizon. Lake Uinta during this period of deposition had reached its maximum extent and occupied much of central and northeastern Utah and northwestern Colorado.

Strata of the uppermost parts of the Green River Formation accumulated during the waning stages of Lake Uinta and include much thin-bedded to platy siltstone, marlstone, and fine- to medium-grained sandstone; oil shale is only a minor component (Untermann and Untermann, 1964; Cashion, 1967). The high salinity and extreme alkalinity of the lake water at times during this period are reflected both in an unusual suite of authigenic minerals, many of them microscopic (Milton, 1957, 1977), and, in some beds, by abundant ellipsoidal solution cavities, from a few centimeters to several tens of centimeters across, that formerly were occupied by nahcolite ($NaHCO_3$). These strata, with their conspicuous "bird's nest" structure, lie 8–45 m below the contact with the Uinta Formation (Untermann and Untermann, 1964).

Uinta Formation

Shrinkage of Lake Uinta during late Eocene time resulted in a gradual, though irregular, increase in fluvial over lacustrine conditions, reflected in the stratigraphic record by the predominantly clastic deposits of the Uinta Formation. Very fine to medium grained, calcitic, medium-bedded to massive sandstone, some of it interbedded with thin- to medium-bedded siltstone, dominates the lower part of the formation. Individual sandstone units as measured by Cashion (1982) in outcrop near Bonanza commonly are 20–60 m thick (fig. 2) and are separated from one another only by thin (0.2–6 m) intervals of marlstone and ash-fall tuff. Most of the sandstone bodies crop out as laterally persistent ledges or steep slopes (Cashion, 1982) and were deposited in a dominantly marginal-lacustrine environment (Franczyk and others, 1989). This sandstone-rich part of the section, more than 400 m thick near the mining camp of Bonanza (fig. 2), hosts most of the thickest gilsonite dikes in the eastern Uinta basin (Pruitt, 1961; Cashion, 1967). Most of the gilsonite mines active today, along five large dikes near Bonanza, exploit gilsonite from this part of the section.

Mixed fluvial and marginal-lacustrine facies characterize the middle part of the Uinta Formation and grade upward into an entirely fluvial sequence (Franczyk and others, 1989) containing much variegated mudstone and channel-form sandstone and minor conglomerate. Sandstone units in the middle and upper parts of the Uinta Formation tend to be much thinner than those lower in the section (see uppermost part of Bonanza section, fig. 2) and are laterally discontinuous, properties ill-suited to the mining of gilsonite dikes within them. The mudstone-rich upper part of the Uinta Formation typically erodes to a badlands topography.

Duchesne River Formation

Overlying the Uinta Formation, and in most places conformable with it, are the late Eocene and Oligocene beds of the Duchesne River Formation. These too are of fluvial origin and consist mostly of weakly cemented channel sandstone, red and maroon mudstone, and some conglomerate derived from source areas to the north in the Uinta Mountains (Untermann and Untermann, 1964; Cashion, 1967; Andersen and Picard, 1972). Beds of the Duchesne River Formation have been eroded from much of the study area but are preserved locally in its northwestern part, where they are host rocks for the upper parts of several gilsonite dikes that at deeper levels were mined.

MAJOR DIKE SYSTEMS OF THE UINTA BASIN

Pruitt (1961) recognized six major geographic systems of gilsonite dikes (plate 1*A*), each separated from neighboring systems by an intervening area wherein the dikes are less abundant or absent, or simply shorter. Crawford and Pruitt (1963) speculated that some of the systems may be interconnected at depth by dikes that fail to reach the surface. Exploration and mapping of gilsonite dikes in some areas is advanced, but in others, particularly where mining had been long restricted by statute, the available information is sketchy at best. Moreover, because most mapping efforts have been directed at dikes of minable width, the number of dikes shown on existing maps probably represents a variable proportion from place to place of those actually present.

In addition to the dikes of the six systems described below, minor dikes are present farther north, near Red Wash; farther south, near the head of Asphalt Canyon; and farther southwest, on Wild Horse Bench (Crawford and Pruitt, 1963). Information presented in the balance of this section is from Pruitt (1961) and Crawford and Pruitt (1963), except where noted otherwise.

Pariette System

The Pariette (Culmer, Toquor) and Castle Peak (Seaboldt, Baxter) dikes are the principal elements of the Pariette system, the westernmost of the six (plate 1*A*). Both dikes crop out in the Uinta Formation, strike about N. 35° W., and measure about 11 km long by only 30–40 cm thick at the surface. At depth, however, they widen to minable thickness; the Pariette dike, for example, measures 90 cm across at the 244-m level and 100 cm across at the 323-m level. The dike has been mined to a depth of 460 m. A third dike of comparable thickness but much shorter length crops out about 3 km northeast of the other two. Also present are smaller dikes of subeconomic dimension, most of these reportedly in a zone about 75 m wide adjacent to the Pariette dike.

The Pariette area is of historic significance as the site of some of the earliest gilsonite mining operations (1890 for the Pariette dike) in the Uinta basin. The area has had a lengthy production history but currently is idle.

Fort Duchesne System

The Fort Duchesne system (plate 1*A*) contains only two prominent dikes, the Carbon (St. Louis) and the Raven (Duchesne). The Carbon dike maintains a thickness of 0.9–1.2 m for about 2.5 km along its 4.8-km length but then gradually diminishes to a few centimeters at either end. It was discovered in 1869 and was the first dike to be mined (about 1888) in the Uinta basin. The nearby Raven dike has an average width of 45–60 cm and a length of about 4.8 km; it has been worked to a depth of 213 m, with indicated reserves extending to at least 900 m (Remington, 1959, cited in Pruitt, 1961). One or the other of these dikes was mined continuously from the late 1800's until about 1946. The Carbon and Raven dikes strike N. 40° W. and N. 37° W., respectively, and crop out in the Duchesne River Formation.

Ouray System

The Ouray system, near the confluence of the White and Green Rivers (plate 1*A*), contains numerous gilsonite dikes of small (fig. 3) to moderate size. Strikes of the dikes are more variable here than in most other areas and range from N. 43° W. to almost east-west; most dikes,

Figure 3. Thin (10 cm) gilsonite dike of Ouray system showing typical appearance of gilsonite dikes in outcrop. Weathered, dull-lustered fragments of gilsonite litter surface. Near Ouray, Utah, SW1/4 sec. 17, T. 9 S., R. 20 E.

however, strike between N. 55° W. and N. 70° W. The longest dike has been traced for almost 18 km but the others less than half that distance; the map of Pruitt (1961) shows lengths of about 1.5–5 km to be characteristic. Almost all of the dikes are 0.6 m thick or less. The one prominent exception, the Pride of Utah dike, is 0.9–1.5 m thick along much of its length. A few dikes have been worked intermittently for gilsonite, locally to a depth of 400 m, but the narrowness of most of the dikes has discouraged production. All of the dikes crop out in the Uinta Formation, the only unit exposed in the area.

Willow Creek System

The Willow Creek system, located south of the Ouray area (plate 1*A*) and perhaps continuous with it, contains two dikes of moderate size plus numerous others of smaller dimension. The two largest dikes strike

Figure 4. Trench along wide, mined-out gilsonite dike of Rainbow system about 0.5 km northwest of abandoned mining camp of Rainbow, SE1/4 sec. 23, T. 11 S., R. 24 E. View southeast. Note wooden stulls that formerly supported planking used as horizontal catwalks by the miners. Trench is about 2.4 m wide.

N. 60°–64° W.; each is 13–16 km long and 0.6–0.9 m wide. Mining has been intermittent. The dikes crop out in the Uinta Formation and may connect at depth to some of the dikes of the Rainbow system (see below) farther east.

Rainbow System

The Rainbow system, in the southeastern part of the Uinta basin (plate 1*A*), contains some of the largest and most productive gilsonite dikes known. At least 16 dikes totaling 116 km in length and having average widths of 0.3 m or more are known from this area. The dikes strike N. 48°–61° W. and crop out from the uppermost beds of the Wasatch Formation to the lower part of the Uinta Formation. Vertical extents of individual dikes range from 90 to 335 m and generally are least in the eastern part of the system, where erosion has cut deeply into the Green River Formation so that only the lower reaches of the dikes are preserved.

Three of the dikes—the Pride-of-the-West, Rainbow, and Black Dragon—are of major dimension. They are approximately in line with one another and are considered by some authors to constitute a single dike totaling almost 39 km in length, making it, on this basis, the largest gilsonite dike known. The northwesternmost segment, the Pride-of-the-West dike, strikes N. 61° W. and is 22.5 km long and about 1.2 m thick. To the southeast it meets the Rainbow dike (fig. 4), which is of slightly different strike (N. 48°–58° W.) (Cashion, 1967), about 7.2 km long, and about 2.5 m thick within a stratigraphic interval from 60 m above to 60 m below the Uinta–Green River contact. When traced southeastward into older beds, however, the Rainbow dike splits into as many as six smaller dikes

Figure 5. Subparallel trenches along mined-out gilsonite dikes of Rainbow system, SW1/4 sec. 30, T. 11 S., R. 25 E. View northwest. Maximum dike thickness is 1.2 m. Prominent "chimney" on skyline is Thimble Rock.

(fig. 5), each about 1 m or less thick (Cashion, 1967); these pinch out downward within the upper part of the Green River Formation. Farther southeast, and at still lower stratigraphic levels, is the Black Dragon dike, stratigraphically one of the deepest dikes in the basin. This dike strikes N. 48° W., crops out for a distance of about 8 km, and is 0.3–1.5 m thick, with pockets as wide as 4 m near the bottom. It cuts strata of the upper part of the Wasatch and lower part of the Green River Formations and pinches out upward within the overlying oil-shale beds. The Black Dragon, Rainbow, and Pride-of-the-West dikes once supported large-scale mining operations, from about the turn of the century to 1936 (when the Uintah Railway ceased operation) and again on a small scale from 1940 to 1955 (Cashion, 1967), but the mines presently are idle.

Cowboy-Bonanza System

The Cowboy-Bonanza system (plate 1*A*), a major concentration of gilsonite dikes, contains 9 or 10 large dikes that collectively have supported mining for a full century, starting about 1890 along the Colorado dike (the only gilsonite dike known to extend into Colorado; hence the name) and continuing today along the Cowboy (Eureka), Bonanza (Independent), Little Bonanza, Little Emma, and Cottonwood dikes. Data given by Pruitt (1961) show that most of the largest dikes exceed 0.9 m in thickness; four are 3 m thick or more along substantial parts of their length. The mammoth Cowboy dike (fig. 6) is the thickest of all known gilsonite dikes at almost 5.5 m; the nearby Bonanza dike is second with a maximum thickness of slightly more than 4.2 m; and the Little Bonanza dike, despite its name, is 4–5 m across in places and averages 2.7 m or more for almost 10 km of its length (Eldridge, 1901; Pruitt, 1961; Cashion, 1967). The combined length of all dikes in this system having a thickness of 0.3 m or more is 84 km. Map lengths of most of the principal dikes are 6 km or more and range upward to about 23 km; moreover, both the Cowboy and the Bonanza dikes reportedly extend much farther northwestward at depth than in outcrop (E.V. Deshayes, quoted in Barb, 1944). Estimated vertical extents of the dikes range from 245 to 425 m (Pruitt, 1961; Cashion, 1967).

Most of the dikes of this system strike N. 55°–65° W. (Cashion, 1967), are widest in the sandstones of the lower

Figure 6. Mined-out trench near southeast end of Cowboy (Eureka) gilsonite dike, SW1/4 sec. 23, T. 9 S., R. 25 E. View southeast. Patches of unmined gilsonite adhere to both walls of trench. Trench is about 3.5 m wide, 60 m long, and 9 m deep. Note lateral jog in dike wall and thin vertical gilsonite seam separated from main part of dike by bedrock (left side of photograph).

part of the Uinta Formation, and taper downward to zero thickness within the underlying oil shales of the upper part of the Green River Formation. A prominent exception is the Colorado dike, which cuts strata of the lower part of the Green River Formation and extends 90 m into beds of the underlying Wasatch Formation. It, like the Black Dragon dike of the Rainbow system, is stratigraphically one of the deepest dikes known in the basin.

PREVIOUS HYPOTHESES ON DIKE ORIGIN AND SOURCES OF GILSONITE

The twin problems of identifying the source beds of gilsonite and explaining the origin of the fractures in which this substance occurs have been long debated. The disparate hypotheses on both topics are reviewed at some length below, for purposes of both scientific importance and historical interest.

1. Stone (1891, p. 155), during one of the earliest investigations of the gilsonite dikes, mentioned that the dikes are known to extend from the Duchesne River Formation downward through the Green River beds "and nobody knows how much deeper." He hypothesized that they extend to "profound depths," where they would intersect "the marine Cretaceous shales" (Mancos Shale), a likely source of petroleum which would then have migrated upward through the fissures. An alternative source recognized by Stone is organic material in the Tertiary lake beds bordering some of the principal exposed dikes then known. Stone favored the latter source and felt that asphalt-saturated strata adjacent to the dikes drained some of their contents into the original fissures, but as to the cause of the fissures themselves he remained silent. Douglass (1928) later offered a similar hypothesis, that the gilsonite was derived from oil shale of the upper Green River Formation and flowed by gravity into the fissures, aided perhaps by (p. 17) "the pressing of the oil out of the shale by the weight of the rocks lying above." Douglass further stated (p. 17) that "we can see no other possible mode of filling of the veins" and (p. 121) "the theory which some have advocated, that the oil residue came up from below under pressure, was denied by all the observed facts." He rejected the notion that the Cretaceous beds were the source of the gilsonite on three grounds. (1) The Mancos Shale is separated from the gilsonite dikes by tens of hundreds of meters of intervening rock, including (p. 122) the "practically impervious clays" of the Wasatch Formation. (2) If fissures through the intervening rocks had allowed the upward migration of fluids, they too should be filled with gilsonite, but no such evidence of vertical connection between the Cretaceous and Tertiary beds had ever been found. (3) (p. 122) "Why look for the source in some mysterious and unknown place beneath when the oiliest formations known could be seen in actual contact with the veins?"

2. Eldridge (1902) clearly felt that the Green River Formation was the ultimate source of gilsonite, citing as evidence (p. 304–305) "the absence of every trace of petroleum in the enclosing sandstones [of the Uinta Formation] and its evident prevalence in the underlying Green River shales." The fissures themselves he felt possibly were due to "the gentle folding" that deformed the basin rocks into their present synclinal form (Eldridge, 1896a, p. 928). If so, he noted, the cracks would have grown from below upward, could extend to considerable depth, and probably would exhibit a tendency to widen downward. The considerable vertical extent of some of the dikes already had been established from field studies. In a later paper, however, Eldridge (1906) abandoned the folding

hypothesis and stated instead that the fractures more likely were due to shrinkage, or contraction, of the strata. He based this changed view on the thick, lengthy dikes of the Cowboy-Bonanza system, which trend at high angles to the strike of the enclosing rocks and thus cannot readily be ascribed to axial fracturing on a growing fold. The causes of the stratal contraction and strong preferred trend of the dikes were frankly acknowledged as unknown. Whatever their origin, Eldridge (1906, p. 444) felt that opening of the fissures would have created a vacuum whose "powerful suction" would have drawn the disseminated bitumen from the shale into the dikes.

A variant of Eldridge's original (1896a) hypothesis was mentioned briefly much later by Untermann and Untermann (1964), who commented (p. 92) that the view most generally accepted at that time was that the gilsonite dikes originated as "tension cracks [that formed] as a result of compaction of the Green River * * * shales and the downwarping of the Uinta Basin syncline which was taking place between the rising Uinta Mountains on the north and the renewed uplift of the Uncompahgre Highlands on the south."

3. Murray (1949, 1950) rejected the notion that disseminated kerogen in the Green River Formation was the source of the gilsonite and favored instead (1950, p. 118) the early formation, during compaction of the beds, of a hydrocarbon derived and "readily eliminated" from the abundant algae in the original sediments, leaving kerogen behind as a residue. In this he presaged some of the modern thoughts on the issue, wherein gilsonite is considered an early expulsion product from the bitumen rather than the kerogen fraction of the organic material in oil shale (Hatcher and others, 1992). Murray also revived an alternative hypothesis, previously rejected by Douglass (1928), that gilsonite had been derived from Cretaceous beds below the dikes. Of the fissures themselves he proposed (1950, p. 118), somewhat enigmatically, that they resulted from "vertical shear" but were [later?] opened by tension. Direct tensile opening of the dikes rather than opening due to hydraulic pressure of the injected gilsonite was favored (1950, p. 118) because of the "lack of extensive impregnation of the sandstone walls with gilsonite and because the cover in places would not have withstood a pressure sufficient to open the fractures by pushing the walls apart." As later discussed, however, the cover in many places did *not* withstand the fluid pressure of the intruded gilsonite but instead was fractured along bedding planes and uplifted as gilsonite sills were injected laterally from the dikes.

4. Crawford (1949) noted the presence of west-trending strike-slip faults in the central Uinta basin south of Duchesne, deduced the sense of slip on one of them as right lateral, and hypothesized that formation of these faults was linked to that of the gilsonite dikes. The dikes, in this view, were regarded as large extension fissures ("tear cracks" in Crawford's terminology) that formed oblique to the faults during post-Eocene compression of the basin. The attendant synclinal downwarping he envisioned to have stretched the underlying beds relative to those higher in the section, thereby (p. 248) "providing cracks which often widen with depth, giving relatively easy access for their filling by heavy viscous fluids from below." Crawford suggested that fault movement was episodic and (p. 251) that sudden slip induced "instant compaction of the bituminous sandstones and marlstones far below," thereby expelling "their fluid and semi-fluid constituents under enormous pressure so that this material immediately would be injected into the open fissures." The position and preferred orientation of the dikes he attributed (p. 249) to "pre-existent lines of weakness in the subjacent basement rocks," evidence for which was claimed not only on the basis of unstated field relations but also on physiographic grounds. The latter refers to the observation that the Uinta River flows roughly parallel to the northernmost gilsonite dikes and thence through a gap in the escarpment along the south flank of the Uinta Mountains, this being taken by Crawford as surface evidence of basement control. The gilsonite itself he considered to have been derived both from the bituminous sandstones of the Wasatch Formation and from kerogen in the Green River beds.

Crawford's hypothesis of fault-related extension fissures was later rejected by Davis (1951, 1957), who felt that evidence of lateral displacement of sufficient magnitude to produce fissures several kilometers long is lacking in the basin.

5. An hypothesis listed (and rejected) by Davis (1951, 1957), but without attribution to source, is that the gilsonite dikes were filled by downward seepage from an overlying asphalt lake or asphaltic source beds. This hypothesis assumes that the fractures that were to become the gilsonite dikes were formed by processes unrelated to dike intrusion and were filled from above rather than below; it also leaves open the question of the original source of the asphalt. One objection to this hypothesis (Cashion, 1967) is that nowhere in the region is there stratigraphic evidence that such an asphalt lake existed, though such evidence could have been lost through erosion. Cashion also mentioned that the geometry of the fractures containing gilsonite is incompatible with this hypothesis but did not detail his reasoning; Davis (1951, p. 38) dismissed it merely on the basis of "the field geology." The principal objection, however, is the abundant evidence for forceful rather than passive intrusion of the gilsonite, as discussed later in this report.

6. An hypothesis frequently mentioned, but whose origin is uncertain, is that the gilsonite dikes formed as fold-parallel fissures along the crestal regions of intrabasin anticlines. Quigley (1950), for example, noted that the trend of the Independent-Tabor vein, if projected southeast, would be approximately coincident with the crest of

the West Douglas Creek anticline and suggested that this led some geologists to propose a link between intrabasin folding and gilsonite-dike formation. Cashion (1967) later noted that most of the veins of the Cowboy-Bonanza and Rainbow systems lie along two northwest-plunging structural noses. The hypothesis probably is an old one, however, because Eldridge (1896a, p. 928) seemed aware of it when he commented that none of the dikes show "the irregularities of fissures formed by the tearing asunder of strata along the axis of an anticline." He further commented that "whereas in some places they [the dikes] are about parallel to the strike of the strata and the main flexures in them, in others they cut the strata diagonally to their strike." Davis (1951, 1957) rejected the anticline hypothesis on the grounds that fissures due to anticlinal folding should narrow downward but numerous gilsonite dikes instead widen with depth. The opposite hypothesis, that the dikes occupy upward-tapering fissures along the troughs of intrabasin synclines, also was mentioned by Davis but with little discussion and no mention of the originator of the hypothesis. Comparison of dike distribution with a recent structure-contour map (Smith, 1981) of the Mahogany bed (fig. 2), within the source-rock interval for the gilsonite, reveals no convincing evidence for either hypothesis.

7. Davis (1951) felt that gilsonite dikes are tensional fissures that formed by differential compaction of the Tertiary sediments above the crests of irregular highs on an erosion surface developed on the underlying rocks. The gradual arcing in strike of the veins, from west-northwest near the Colorado State line to northwest farther into the basin, was regarded by Davis as evidence of possible subsurface control on dike orientation and position. In a later (1957) publication he credited this idea to G.H. Hansen of Brigham Young University and modified it by invoking a combination of differential compaction and (p. 156) "structural pressures exerted on the Uinta Basin when it was uplifted * * * beginning during the early Tertiary period and continuing to the present time." No further evidence to substantiate his claims is given in either paper.

8. Hunt (1963), like others before him, viewed the gilsonite dikes as "tensional cracks" but related the cause of tension to regional uplift and the removal of 600–1,800 m of overburden rather than to basin downwarping, folding, or faulting. He felt that the cracks opened slowly as uplift progressed and were simultaneously filled with the bituminous material that would later, upon increasing polymerization, harden into gilsonite. Hunt, like Eldridge (1906), suggested that opening of the cracks reduced the pressure within them, thereby favoring movement of the viscous bitumens from the rock into the growing voids.

9. Cashion (1967), in his study of the hydrocarbon resources of the Green River Formation in the area of the Cowboy-Bonanza and Rainbow dike systems, reviewed existing hypotheses on the origin of gilsonite. In addition to the sources listed above, Cashion mentioned the opinion of some geologists that gilsonite was derived from oil in the Weber Sandstone, a unit of Pennsylvanian and Permian age that in some nearby fields is a prolific producer of petroleum. The large vertical distance between the dikes and the Weber Sandstone, and especially the dissimilar trace-element suites between crude oil from this unit and gilsonite, probably are sufficient to invalidate this hypothesis (Cashion, 1967, p. 35). In addition, observation that no known dike cuts all the way through the Mahogany zone of the upper part of the Green River Formation, but that most pinch out above or within it, seemingly argues against derivation of gilsonite from any source stratigraphically below this zone except for those few dikes that cut the lower part of the Green River and the Wasatch Formations. The absence of any observed conduits that could have served to convey fluids to the gilsonite dikes from stratigraphic levels below them has repeatedly been invoked as an objection to deep sources of gilsonite since the days of Douglass (1928).

Cashion (1967) also commented on one remaining hypothesis, unidentified as to source, that gilsonite was derived from sandstones of the lower part of the Uinta Formation, in which many dikes attain their maximum thickness. Eldridge (1902), however, had earlier mentioned the principal objection to this hypothesis, that these sandstones contain no evidence of petroleum residue except immediately adjacent to the dikes themselves, where gilsonite-impregnated sandstone is evidence for infiltration from the dike into the sandstone rather than the converse. To this Cashion added another objection, that gilsonite probably was emplaced as a highly viscous fluid, and no mechanism for the almost complete removal of such a fluid from sandstone has ever been demonstrated or proposed.

The various hypotheses listed above are nothing if not diverse. Speculation on the ultimate source of the bitumen now found in the gilsonite dikes includes mention of every stratigraphic unit adjacent to the dikes and several of those below, as well as an asphalt lake above. Hypothesized source rocks range in geologic age from Pennsylvanian to middle Eocene and span a stratigraphic interval of more than 4,000 m. Formation of the dikes has been related by different authors to basin downwarping, fracturing along the crests of intrabasin anticlines or the troughs of intrabasin synclines, contraction of the host rocks, strike-slip faulting in the basin, differential compaction over an irregular erosion surface, and erosional unloading resulting from regional uplift. Intrusion of the gilsonite has been described as slow by some authors and almost instantaneous by others, under pressures that range from "enormous" through near zero (gravity flow into open cracks) to negative (suction of fluids into a crack suddenly opened). And finally, the various hypotheses range from well-reasoned consideration of the available evidence to wholesale and almost baseless speculation couched in vague language—altogether, a fascinating body of literature spanning almost a century of work.

The long-debated question of the ultimate source of gilsonite was at last resolved through the comprehensive work of Hunt and others (1954) and Hunt (1963), who compared the physical and chemical properties of gilsonite to those of bitumens still disseminated in hypothesized source beds. Infrared spectra of the disseminated bitumens show a clear and progressive change upsection through the Wasatch, Green River, and Uinta Formations, enabling Hunt and others (1954) and Hunt (1963) to identify the source beds of gilsonite and other vein bitumens in the basin by comparing their spectra to those of the disseminated material from different horizons. The results showed convincingly that "gilsonite is quite similar to the bitumen disseminated in the upper Green River oil shale and differs from those both below and above it. This interval * * * contains the Mahogany ledge [fig. 2] and is generally more bituminous than other sections in the basin" (Hunt, 1963, p. 271). Hunt suggested that the most important factor governing the composition of the bitumens was salinity of the lake in which the original sediments were deposited. The changing and generally increasing salinity with time correlates stratigraphically with (1) the change from precipitated calcite in the basal strata to highly soluble Mg-Fe-Na carbonate minerals in the younger strata; (2) an upward decrease in the abundance of macroscopic fossils of benthic organisms; and (3) a systematic upward change in the molecular structure of the associated bitumens. Depositional environment thus exerted the ultimate control on the character of the later vein deposits. Subsequent work, including evaluation of biomarker composition by gas chromatography and mass spectrometry, has reaffirmed the close relationship between gilsonite and oil shale of the Green River Formation (Khavari-Khorasani, 1974; Palacas and others, 1989; Anders and others, 1992; Hatcher and others, 1992).

STRUCTURE OF GILSONITE DIKES

Dimensions

The gilsonite dikes of the Uinta basin range in size over several orders of magnitude, from small dikelets at the outcrop scale to thick dikes traceable for many kilometers across the countryside and having dimensions comparable to those of the basaltic dikes of Hawaii or Iceland. The longest dike known, assuming that its three segments form one continuous dike in the subsurface, is the combined Pride-of-the-West/Rainbow/Black Dragon dike, traceable in outcrop for a distance of about 39 km. Among individual dikes traceable in continuity at the surface, the Pride-of-the-West and Cowboy dikes have been followed for almost 23 km and several other dikes for more than 15 km. Most, however, are much shorter, and more than half of all mapped dikes are less than 5 km long, at least at the surface.

Individual dikes are several millimeters to 5.5 m thick; 0.5 m is the approximate minimum thickness for mining. Only those at least 0.3 m thick, however, have been well explored, and existing maps probably underrepresent the number of dikes of subeconomic thickness. The thin dikes, too, are those most likely to be concealed by surface debris; their presence in some areas became apparent only when shallow trenches were cut by bulldozer to follow dike extensions.

The probable vertical extents of the gilsonite dikes in any given area can be estimated from the map of Smith (1981), which shows depths to the Mahogany bed (fig. 2), the richest oil-shale bed within the source interval for gilsonite. These depths increase markedly from southeast to northwest and range from several tens of meters for the southeasternmost dikes of the Cowboy-Bonanza and Rainbow systems, near the basin margin, to 1,500 m or more for dikes of the Fort Duchesne system just south of the structural axis of the basin. The map should be used with caution, however, because some dikes appear to have propagated great distances laterally and thus may not be connected to the oil-shale beds directly beneath, whereas others (Black Dragon, Colorado) are known to extend stratigraphically *below* the oil-shale interval and into the underlying Wasatch Formation. The same cautions apply to our later discussion on depths of emplacement.

Minimum vertical extents of some dikes are known from mining and exploratory drilling. Some dikes of the Pariette and Ouray systems, for example, have been mined to depths of 305–455 m, and indicated reserves of the Raven dike of the Fort Duchesne system extend to at least 900 m below the surface (Remington, 1959, cited in Pruitt, 1961). Other dikes in areas of appreciable topographic relief can be shown in outcrop to cut 200–425 m of section; Cashion (1967) discussed several examples from the Cowboy-Bonanza and Rainbow systems. From such observations it is apparent that dikes of minable thickness tend to be vertically extensive as well; indeed, Crawford and Pruitt (1963) stated as fact that many dikes in the western part of the area extend downward 900 m or more.

Associated with the major dikes are many others of subeconomic thickness (fig. 7) and more limited vertical extent. The low end of the size spectrum is represented by dikelets that cut only a few centimeters to several meters of section and that originate from or terminate against gilsonite sills. The significance of such dike-sill networks to the mechanics of intrusion is discussed in a later section.

Cross-Sectional Shape

Most gilsonite dikes in gross aspect are thin, tabular intrusions whose vertical and horizontal dimensions vastly

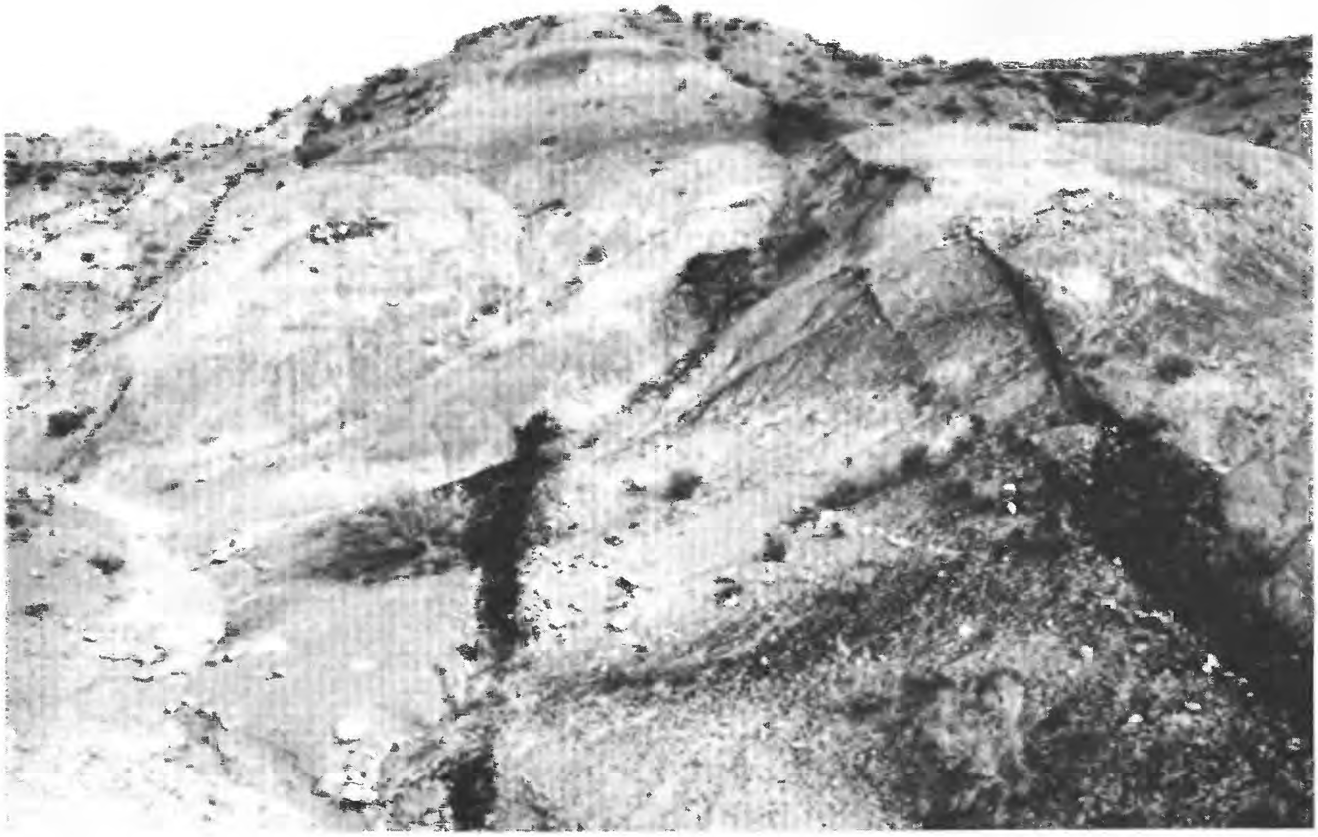

Figure 7. Unmined gilsonite dikes near northwest end of Cowboy dike, SE¼ sec. 32, T. 8 S., R. 24 E. View southeast. Dikes locally branch, pinch out, and are offset from each other in plan view. Note that individual dikes, although subparallel in gross aspect, curve in strike over short lateral distances.

exceed their thickness. Few dikes are more than 3 m thick, but the known and suspected vertical extents of most of them exceed 300 m except where erosion has cut deeply into them, as along some dikes of the Rainbow system. The lengths of numerous dikes exceed their thickness by three orders of magnitude or more. Eldridge (1906) commented on one dike so thin it appeared like a crayon mark across the landscape yet was traceable in continuity for 120 m; its length exceeds its thickness by a factor of more than 9,500.

Many dikes are thickest in the thick sequence of well-cemented sandstones and sandy siltstones of the uppermost part of the Green River Formation and lower part of the Uinta Formation (Douglass, 1928; Pruitt, 1961; Cashion, 1967) and then thin upward and pinch out within the more variably cemented channel sandstone bodies and enclosing mudstone of the middle and upper Uinta and overlying Duchesne River Formations. The dikes also thin downward, below the thick sandstone interval, and display a pronounced tendency to bifurcate into multiple thin dikelets as they pass downward into the bituminous marlstones (oil shales) of the upper Green River Formation (Eldridge, 1901; Pruitt, 1961). The downward termination of one large dike, the Cowboy dike, early was illustrated (fig. 8) and described by Eldridge (1901). In the sandstone the gilsonite forms a single, tabular dike about 3 m thick, but this upon entering the underlying marlstone splits into a score of minor dikelets extending 15–90 m below, occupying a zone 12–15 m wide at its widest and then narrowing downward as the lowermost dikelets wedge out about 45 m, according to Cashion (1967), above the Mahogany bed. Those few dikes known from below the rich oil-shale sequence wedge out *upward* in similar manner; the upper part of one such dike only a few meters below the Mahogany bed is shown in figure 9. These dikes, too, are widest in sandstone and pinch out downward within the shale and thinner sandstone beds of the lower Green River and upper Wasatch Formations (Cashion, 1967). No dikes from either above or below the Mahogany oil-shale zone are known to cut completely through that zone, though numerous dikes terminate either within or near it (Cashion, 1967, fig. 9). That many dikes

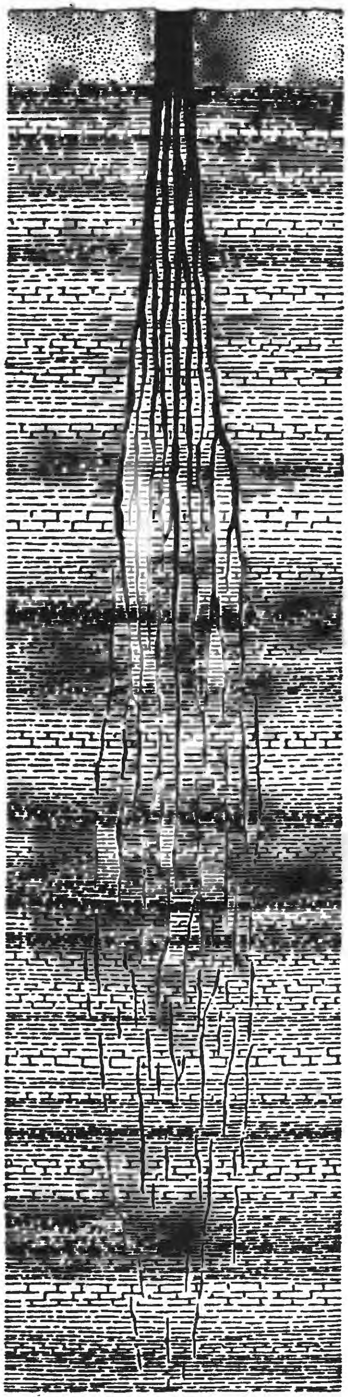

Figure 8. Downward splitting and termination of southeast end of Cowboy dike within upper part of the Green River Formation, as sketched by Eldridge (1901). Probable location is SW1/4 sec. 23, T. 9 S., R. 25 E., along the south-facing cliffs bordering the canyon of the White River near Little Dicks Bottom, 200 m or less from trench along same dike shown in figure 6.

pinch out within this stratigraphic interval is attributable to the nonbrittle nature of bitumen-rich oil shale, which readily deforms by distributed intergranular strain and consequently is difficult to fracture.

The generalized description of gilsonite dikes as being thickest in the thick sandstone-rich interval near the Uinta–Green River Formation transition and gradually thinning above and below that interval was described by Pruitt (1961) as an oversimplification, but basically correct. For that reason, gilsonite dikes in the eastern part of the basin, where they crop out in the upper Green River and lower Uinta Formations, commonly are of minable thickness but generally are expected to narrow and split with depth. Dikes in the central part of the basin, in contrast, crop out in the upper Uinta and Duchesne River Formations and commonly are narrow at the surface but are expected, as a general proposition (Pruitt, 1961), to widen with depth. The Pariette dike, 30–40 cm thick at the

Figure 9. Upward pinchout of Black Dragon dike near its northwestern end, a few meters below Mahogany bed of the Green River Formation. Note abundant fragments of wallrock included in gilsonite. 200 m northwest of Black Virgin mine (abandoned), NW1/4 sec. 4, T. 12 S., R. 25 E. View northwest.

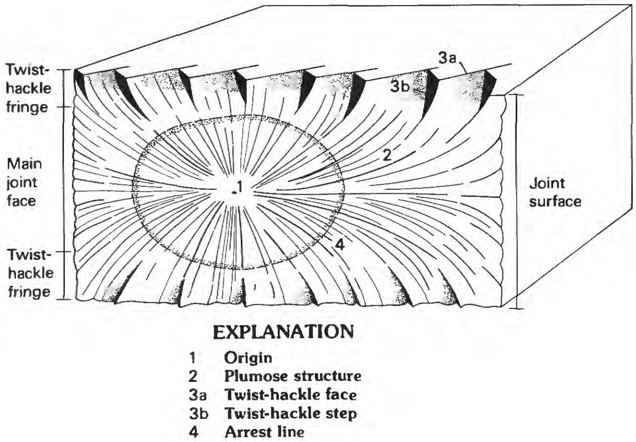

Figure 10. Idealized fracture surface showing common fracture-surface structures and terminology used in this report. Fractures begin growth at the origin point and propagate outward, parallel to the plumose structure. The orientation of surface structures on dike walls thus can be used to infer directions of dike propagation. Modified from Barton (1983).

surface but 100 cm across about 300 m below, is a prominent example.

Morphology of Dike Walls

The walls of gilsonite dikes, though commonly described as showing little irregularity (Eldridge, 1896a; Murray, 1949; Cashion, 1957; Davis, 1957; Henderson, 1957), in detail are actually quite complex and of different character from one rock type to another. All of them display features indicative of extensile failure of the rock (see figure 10 for terminology). We next describe properties of the dike walls for three common rock types within the stratigraphic interval most commonly mined, from the uppermost part of the Green River Formation through the overlying Uinta Formation. Dike walls in deeper and shallower strata were examined only briefly but correspond in their general characteristics to those described here.

Sandstone and Siltstone of the Lower Uinta and Uppermost Green River Formations

Mining of gilsonite from the sandstone-rich lower part of the Uinta Formation has taken place for more than a century, most notably and recently along dikes of the Rainbow and Cowboy-Bonanza systems. As noted previously, individual sandstone bodies low in the formation tend to be laterally persistent, very fine to medium grained, calcitic, firmly indurated, and medium bedded to massive; some also contain much thin-bedded siltstone. Additional mining has taken place within underlying, thin-bedded siltstone and subordinate sandstone beds of the uppermost Green River Formation, which represent a transitional lithology to the Uinta beds. Open trenches along mined-out gilsonite dikes in these rocks afford numerous opportunities for study of dike-wall structure, about which surprisingly little has been said in the geologic literature. One such trench, a readily accessible narrow slot about 3.5 m wide, 60 m long, and 9 m deep (fig. 6), is near the southeast end of the Cowboy (Eureka) dike in the upper Green River Formation. The dike at this locality has an average strike of N. 56° W. and dips within 1° of vertical. The wallrock in the trench is a thin- to medium-bedded, nonporous, well-cemented, slightly dolomitic siltstone to very fine grained sandstone, an ideal lithology for studying the fine structure of the fractures that collectively form the walls of this very large dike. From a distance the walls appear relatively smooth, parallel, and continuous, with only minor undulations along strike, much as they have

Figure 11. Southwest wall of Cowboy dike in mined-out trench near southeast end of dike. Narrow, steeply dipping sunlit fractures are step faces between broad twist-hackle faces of the large, composite extension fracture that forms dike wall. Same locality as figure 6 but at opposite end of trench and viewed from opposite direction.

been described for gilsonite dikes in general by earlier workers. On close inspection, however, all impression of structural simplicity disappears; each dike wall is complexly stepped and composed of multiple, overlapping fractures (fig. 11). Remnant patches of gilsonite still adhere to many of the surfaces. Observed characteristics of the dike walls in this area include the following.

1. The fractures that collectively form the dike walls are large. Most can be followed along strike for 15–35 m before passing behind another fracture that overlaps it, and all have vertical extents greater than the depth of the trench (8–10 m). These are minimum dimensions only; no fracture could be traced along its full extent.

2. The fractures are almost invariant in dip but locally undulate gently along strike. Dike walls along the most pronounced undulations deviate as much as 15° from the overall dike trend within a lateral distance of several meters, though typically the departure is much less.

3. Lateral jogs in the dike walls (fig. 6) are present in a few places.

4. Each fracture displays prominent twist hackle (fig. 11). The distance between adjacent twist-hackle faces commonly ranges from 6 to 15 cm.

5. Step faces that strike almost perpendicular to the trench and that dip 50°–70° SE. (fig. 11) connect adjacent twist-hackle faces at many points. The step faces are irregular, narrow, upward-tapering fractures that, collectively with the abundant twist-hackle surfaces, impart to the dike walls considerable local relief.

6. Plumose structure is a constant feature of the fracture surfaces but is readily visible only during the morning hours, when the sun is at a low angle to the dike walls. The plume lines—actually a series of discontinuous, almost microscopic, tapering ridges developed in great profusion—plunge at angles of 50°–70° SE., roughly parallel to the irregular step faces.

7. Large arrest lines were observed on two fracture surfaces. Also noted near the midpoint of the trench was a well-developed hook of one fracture into another.

8. No evidence of shear, in the form of slickensides or stratal offset between dike walls, was seen in the trench.

Similar relations were found within the Green River and Uinta beds at other places along this lengthy dike: within additional trenches 1.5 km to the west-northwest and within the E–29 mine of the American Gilsonite Company at Bonanza. In each place the fractures display the full complement of surface features—plumose structure, twist hackle, arrest lines, and hooks of one fracture into another—characteristic of extensile failure of fine-grained, well-cemented rock, and in each place evidence of shear along the dike walls is lacking. The arrangement of surface structures on the dike walls in the southeasternmost trench (fig. 11) corresponds to the lower left part of the idealized fracture surface depicted in figure 10 and indicates that this part of the dike propagated to the southeast and *downward*, consistent with the location of the trench near the southeasternmost known outcrop of the dike, and very near its base. Thus, the gilsonite within the dike probably was derived from a source to the northwest, not from the bituminous marlstone directly beneath. The surface structures also imply that the fractures as presently exposed are but small portions of much larger individuals whose full size remains unknown. From these relations we conclude that the Cowboy dike formed through dilation, perpendicular to the dike walls, of an originally narrow zone of large, overlapping, and partially interconnected extension fractures. The full width of this zone is unknown due to debris cover at the top of the trench, but the main, intruded part of the fracture zone probably was only 1–2 m wide before dilation. Gilsonite filled the fracture zone to a thickness of 3.6 m and also penetrated the dike walls along individual twist-hackle faces, many of which are occupied by thin (1–4 cm, but locally as thick as 20 cm), tapering, wedgelike deposits of gilsonite that form multiple minor offshoots at acute angles from the main dike. These, as well as lateral jogs of the dike wall that resulted from the filling of parallel but noncoplanar fractures (fig. 6), are products of the original fracture-zone geometry.

Channel-Form Sandstones of the Middle and Upper Parts of the Uinta Formation

Dike walls in thickly bedded to massive, medium- to coarse-grained and locally conglomeratic channel sandstone bodies of the Uinta Formation tend to be of simpler morphology than those in thinner beds of finer grain. Relative to stratigraphically lower, marginal-lacustrine beds, the commonly coarser grain size, poor sorting, and generally weaker induration of the fluvial channel sandstones are properties poorly suited to the development of small-scale fracture-surface structures; plumose structure, in particular, commonly is absent or only faintly developed as a result. The massive to irregularly bedded nature of much of the sandstone likewise suppressed formation of twist hackle, which elsewhere, in more well bedded successions of strata, most commonly is present along the interface between two beds of dissimilar lithology. Dikes in the massive channel sandstones thus commonly lack some of the morphologic complexity of dikes deeper in the section. The western end of the Cowboy dike, for example, appears from mined-out trenches in discontinuous sandstone lenses to have fairly smooth, low-relief, sinuous walls of relatively simple structure, quite different from those described above along the same dike in deeper strata. Doubtless it is this and similar dikes that contributed to the common literature descriptions of gilsonite dike walls as being relatively smooth and featureless.

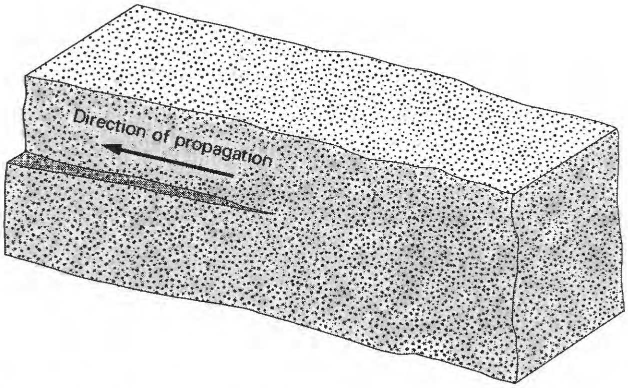

Figure 12. Splitting of dike wall along bedding surface. Split sections of dike wall converge in direction opposite that of overall dike propagation.

One additional feature of gilsonite dikes in the Uinta channel sandstones is deserving of note. Although many of the sandstones are massive, some contain local, discontinuous stringers of finer grained material. Dikes propagating laterally commonly split in two upon encountering such stringers, with the fractures above and below then following similar but noncoincident courses (fig. 12). In this manner large, vertically extensive dike walls locally split into two or more long, ribbonlike segments, each ending abruptly upward or downward against solid rock. In local parlance these are called offsets and caused much consternation during mining (Henderson, 1957), but it must be emphasized that the geometry is a natural result of fracture propagation through inhomogeneous rock and not the result of later segmentation of a once continuous dike by faulting.

Mudstone of the Upper Part of the Uinta Formation

Mudstone in many places is the dominant lithologic component of the upper, fluvial part of the Uinta Formation. Much of the mudstone is only moderately indurated and erodes to rounded, debris-covered slopes offering few good exposures of gilsonite dikes. Locally, however, manmade exposures and protected areas beneath resistant sandstone ledges reveal good examples of dike-wall geometries in the fine-grained beds. Individual dikes in mudstone commonly exhibit curved, irregular, nonmatching walls and thus pinch and swell, locally dramatically, along their length. The mismatched walls are a primary feature and not the result of tangential offset, the absence of which can be demonstrated in many places. Within individual mudstone beds, for example, removal of the gilsonite and closing of the dike would not restore the bed to a continuous, unbroken sheet; the two dike walls cannot be matched no matter how one is positioned relative to the other. Distortion of the wallrock to accommodate varying dike width is thus implied. Bulbous apophyses of gilsonite (fig. 13), noted in several places, also imply significant distortion of the enclosing mudstone. The overall geometry of these dikes is reminiscent of that of clastic dikes intruded into stiff but incompletely lithified mud, examples of which have been described from numerous localities (Dzulynski and Radomski, 1956; Hayashi, 1966; Kholodov, 1978, among others). We infer a similar

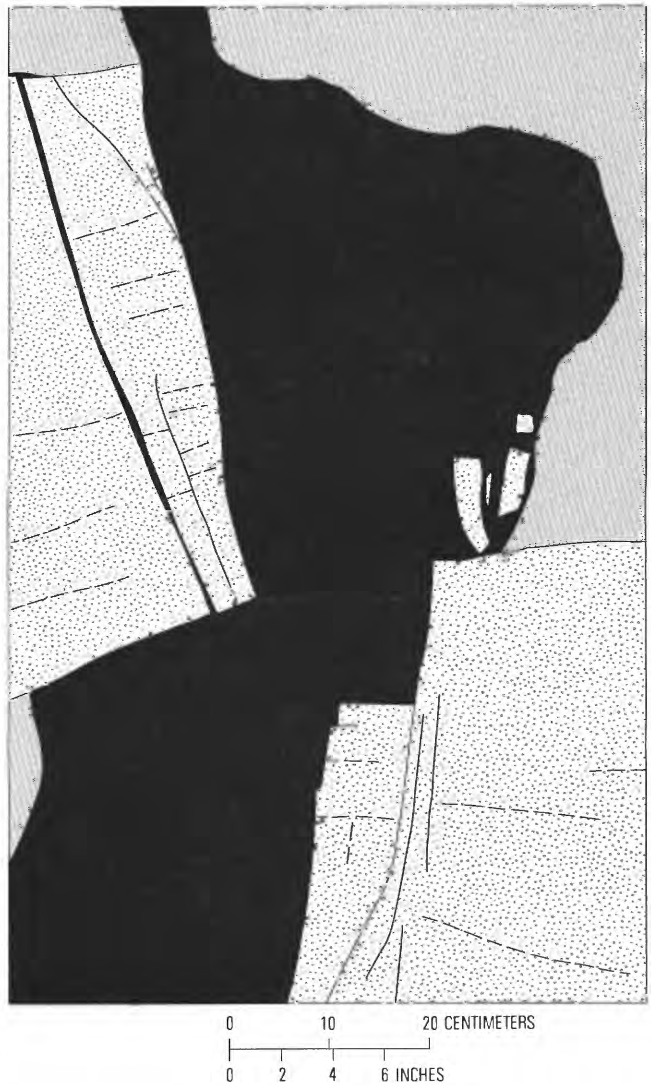

Figure 13. Vertical section through bulbous mass of gilsonite (black) at offset contact between thin (0.5 m) sandstone bed and mudstone of the Uinta Formation, within an exploration pit along dike of Pariette system near Castle Peak in NW¼ sec. 4, T. 9 S., R. 17 E. The gilsonite dike, upon encountering the resistant sandstone bed (dotted pattern), first bowed it upward and then broke through, offsetting the bed about 0.3 m (left side higher). The southwest wall of the dike, where it abuts the sandstone (upper left), is almost planar and defined by a dike-parallel fracture, but to the northeast (upper right) the dike bulged outward within deformable mudstone (shaded areas) to form a prominent protrusion. Traced from photograph. Modified from Verbeek and Grout (1992, fig. 6).

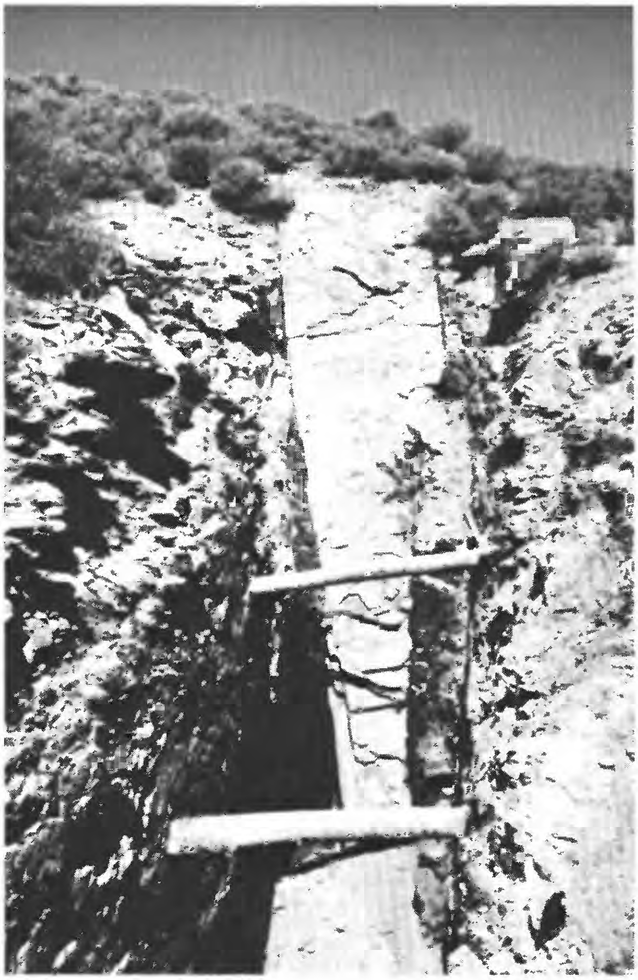

Figure 14. Large, detached and rotated slab of wallrock (center of photograph, facing observer) embedded in gilsonite at northwest end of trench along mined dike of Rainbow system, SW 1/4 sec. 30, T. 11 S., R. 25 E. Slab is tilted about 60°; trench is about 1 m wide.

condition for the Uinta mudstones, which during dike formation clearly were capable of extensile failure yet were still sufficiently weak to distort plastically upon forceful intrusion. The semiductile nature of the Uinta mudstones contrasts strongly with the uniformly brittle response of the associated sandstones and of deeper strata.

Individual fractures forming dike walls in mudstone commonly are 2–10 m long, significantly shorter on average than their counterparts in sandstone; the change in fracture size commonly is visually apparent on a local scale where individual dikes can be traced from mudstone through sandstone and back into mudstone. Hooking of one fracture into an adjacent one is common, as is the repetitive splitting of one sinuous dike segment into two and locally their rejoining to enclose lens-shaped masses of rock. None of these features is common in the associated channel sandstones, where instead much longer, single dikes of simpler morphology and greater thickness are the rule.

Characteristics of Dike Interiors

Gilsonite in most of the dikes examined is of almost homogeneous appearance and contains almost no foreign matter other than local concentrations of rock debris dislodged from the dike walls. Detrital sand grains and microscopic particles of authigenic quartz and barite (Monson and Parnell, 1992) are additional but volumetrically insignificant contaminants. Properties conventionally used in igneous dikes to infer flow directions and sequence of emplacement, such as aligned phenocrysts and variations in grain size, color, and mineralogic composition from dike walls to dike interior, are frustratingly absent in gilsonite dikes. Examination of gilsonite by reflected light, scanning electron microscopy, and electron probe techniques likewise revealed no evidence of flow texture at the microscopic level (Monson and Parnell, 1992). Smooth-walled, hollow ellipsoidal cavities 1–3 mm in diameter, resembling vesicles in a lava flow, were noted in a few places and offer limited information on flow trajectories and the physical properties of the gilsonite during emplacement, but analogous cavities in other places are almost spherical and seem to have formed after flow ceased. Cavities containing water were sometimes found during mining (Cashion, 1967).

Several workers have reported occurrences of different types of gilsonite in the same dike but differ on how these should be interpreted. Lenses of lustrous, homogeneous "selects" ore embedded in duller lustered and commonly minutely fractured "seconds" ore have been taken by some (Douglass, 1928; Murray, 1949; Davis, 1957) as evidence of multiple stages of intrusion; such lenses are present, for example, in the Little Bonanza dike "to great depth" (Pruitt, 1961, p. 35). Others, however, argue that these are purely secondary effects and that no differences in original composition or physical properties are implied by the variations in appearance (Pruitt, 1961). In a later section we present new evidence that some of these effects are indeed secondary but that episodic intrusion nevertheless is consistent with the postulated evolution of the dikes.

Angular fragments of wallrock embedded in gilsonite are known from many dikes (Eldridge, 1896a, Maguire, 1900; Douglass, 1928; Henderson, 1957; Pruitt, 1961; Kilborn, 1964). The clasts vary greatly in size, from small chips less than a centimeter across (fig. 9) to large slabs more than 5 m long (fig. 14); the latter, however, are uncommon. Blocks and slabs of small to moderate size, with two faces defined by (originally) dike-parallel fractures and two others, perpendicular to them, by bedding, are locally abundant within dikes cutting the well-laminated beds of the Green River Formation (fig. 15). Large,

Figure 15. Clasts of marlstone wallrock detached from dike wall and embedded in gilsonite of Black Dragon dike. Portal to small adit near northwest end of dike, about 100 m northwest of Black Virgin mine (abandoned), NW1/4 sec. 4, T. 12 S., R. 25 E.

flat to slightly curved plates wedged or spalled from dike walls locally are present higher in the section, in the sandstones of the Uinta Formation. In places such slabs have rotated away but not separated completely from the dike wall, leaving a wedge of gilsonite between.

Rock debris within dikes is irregularly distributed, being absent or sparse in most places but locally abundant where wallrock lithology was conducive to spalling and where downward movement of debris was impeded either by narrowing dike width or by flat ledges where dike walls are "offset" along bedding. Pruitt (1961), for example, noted that the lower reaches of the Cowboy vein were in places so clogged with debris as to render mining impossible and that the bottom part of the Black Dragon vein was considered unworkable for the same reason. At Bonanza, where drifts along the bottoms of veins were driven to serve as ore passes, embedded rock was also frequently encountered during mining (Kilborn, 1964). The tendency for rock debris to accumulate in the lower reaches of dikes, particularly near the source beds where wide dikes tend to split downward into multiple thinner ones, was noted as long ago as 1901 by G.H. Eldridge. Most such debris, however, probably has not sunk great distances through the dikes but instead was derived from nearby wallrocks particularly susceptible to spalling. The abundance of debris in dikes near their source beds in the middle and upper parts of the Green River Formation, for example, probably owes more to the close network of dike-parallel fractures and to the ready splitting of the well-laminated rock along bedding than to a shower of clasts from above. Local derivation can be demonstrated in many places; Eldridge (1906), for example, commented that inclusions of wallrock often can be matched to their points of detachment from dike walls, and Henderson (1957) noted that large rock inclusions in mined dikes at Bonanza generally could be matched to a corresponding cavity in the dike wall not more than 15 m higher. Similarly, the rock fragments shown in figures 9 and 15 have not moved much from their original stratigraphic positions, and no one, to our knowledge, has matched individual clasts within any dike to strata far above. The apparently limited movement of most clasts through the dikes, despite the large density contrast (1.3–1.7 g/cm^3) between gilsonite and the clasts, implies that the gilsonite was emplaced as a highly viscous fluid (Henderson, 1957; Cashion, 1967).

Dike and Sill Networks in Three Dimensions

The continuous, rectilinear to gently curving traces of gilsonite dikes as portrayed on small-scale maps (plate 1*A*) offer scant indication of the structural complexity of these intrusions. Some dikes are indeed continuous or almost so for lengthy distances, but others—especially those cutting fine-grained beds, as alluded to previously—are discontinuous, segmented intrusions of elaborate three-dimensional structure. Particularly instructive examples are present among the dikes of the Ouray system, about 3 km south of the White River, where abandoned mines and exploration trenches provide good exposures. Several of the dikes were examined in detail where they cut variably cemented mudstone and sandstone, the latter in discontinuous bodies of irregular shape and dimension, of the upper part of the Uinta Formation. Tracing the dikes laterally revealed many places where they abruptly step to the left or right, thicken or thin dramatically, or split into discontinuous, subparallel segments having multiple branches and cross dikes. Figures 16 and 17 show the geometry of one such dike in plan view and vertical section, respectively, where the dike narrows in mudstone to subeconomic thickness. Both figures were prepared by first establishing a meter-square grid over the face to be mapped and then taking all measurements to an estimated accuracy of 1 cm or better

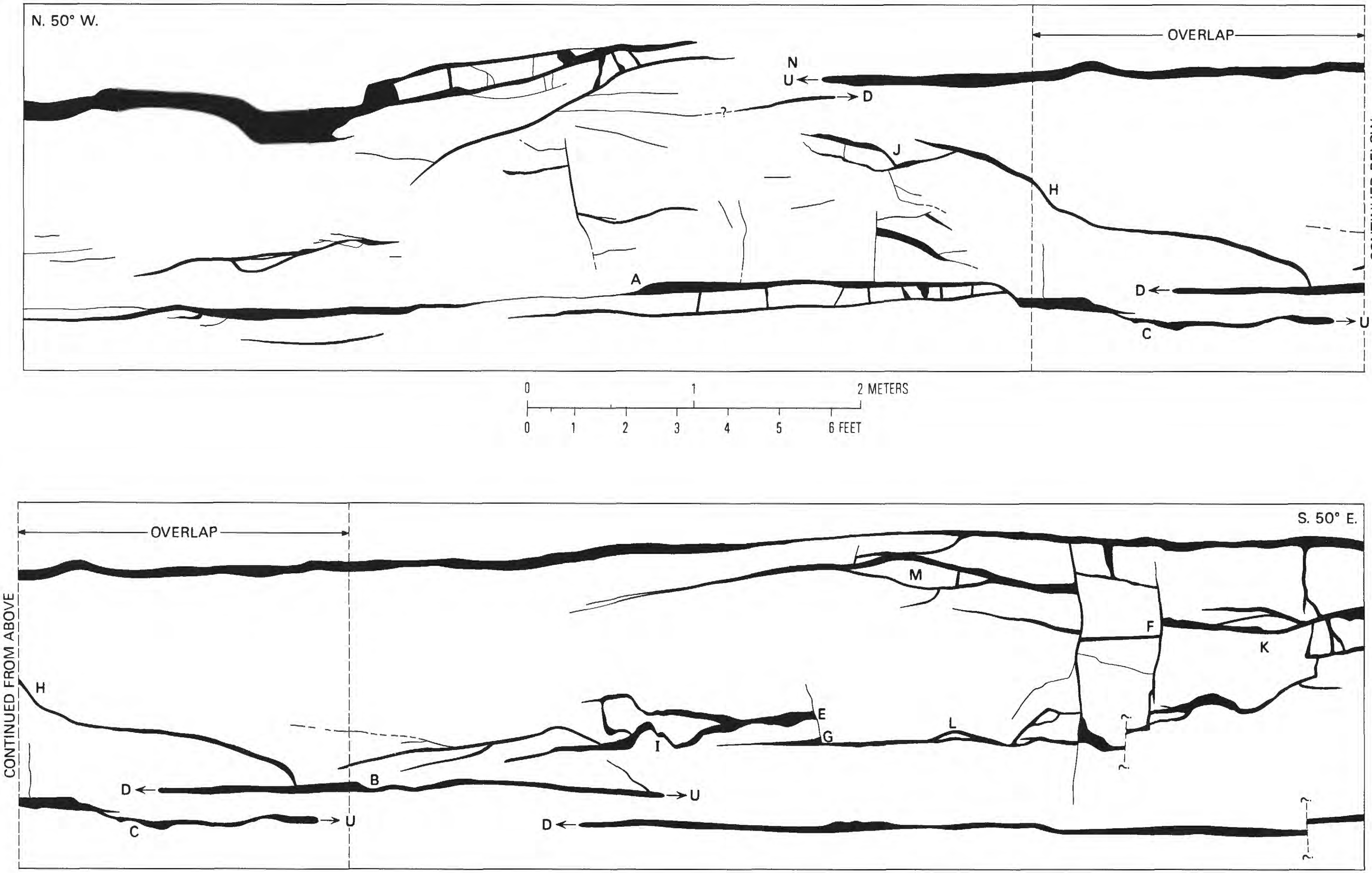

Figure 16. Floor of exploratory bulldozer trench along typical complex dike of Ouray system in mudstone of upper part of Uinta Formation. Note discontinuous, branching nature of dike segments, prominent pinch and swell along individual segments, and numerous cross segments. Letters refer to features discussed in text. SW1/4 sec. 17, T. 9 S., R. 20 E., about 6.5 km south-southwest of the town of Ouray. Modified from Verbeek and Grout (1992, fig. 7).

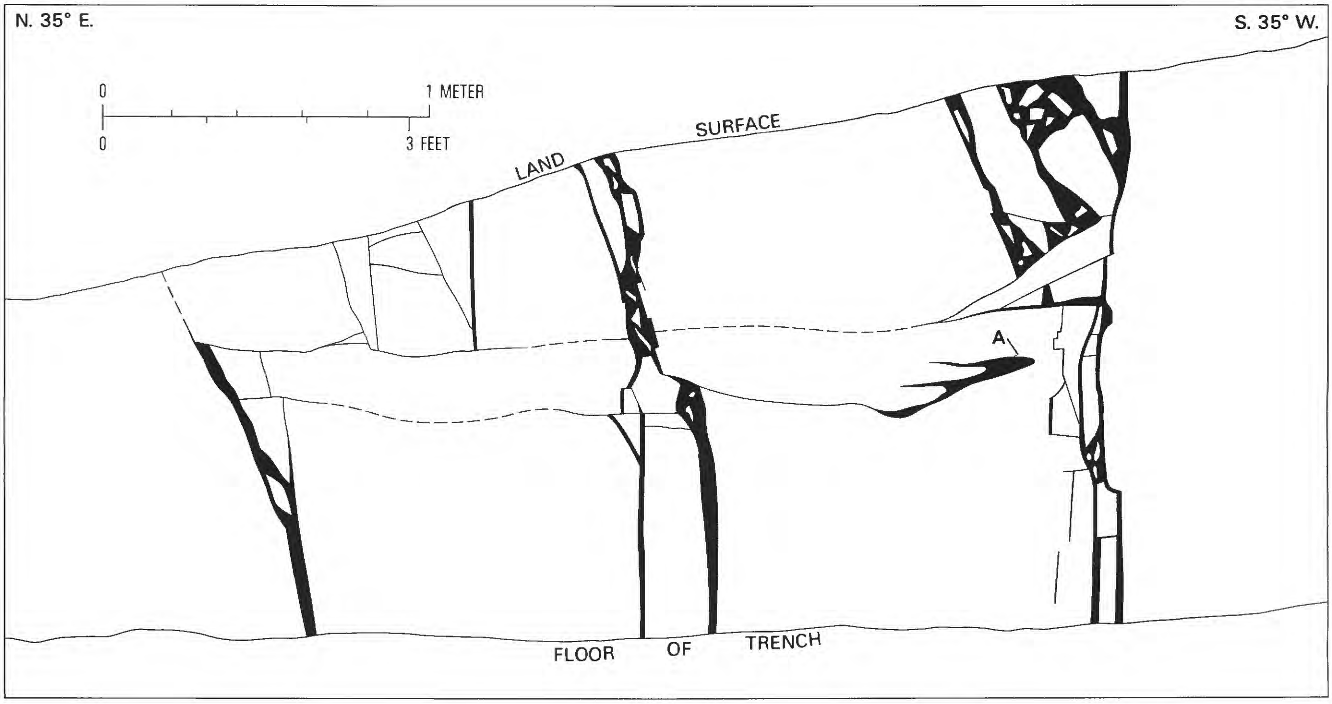

Figure 17. Vertical section showing geometry of typical dike of Ouray system in mudstone of upper part of the Uinta Formation. View southeast of wall of shallow exploratory trench cut across dike (same dike as fig. 16). Note numerous sills and dikelets, variability of dike-sill geometry, and common occurrence of small fragments of wallrock detached from dike walls and embedded in gilsonite. Dashed lines indicate fractures devoid of gilsonite but continuous with sills. NE¼ sec. 20, T. 9 S., R. 20 E. Modified from Verbeek and Grout (1992, fig. 8).

with a steel tape. The results were plotted in the field on quadrille paper and extensively checked so that details of dike form and interconnection could be faithfully depicted.

Noteworthy aspects of dike geometry, as exposed over an area of 45 m^2 on the almost horizontal floor of a shallow bulldozer trench (fig. 16), include the following.

1. Individual dike segments have sinuous, nonmatching walls and in numerous places (such as at points A, B, and C) show abrupt changes in thickness over short lateral distances. Both characteristics, as mentioned previously, indicate that the mudstone was not yet fully lithified when intrusion occurred.

2. Vertical dike segments are of two main types: longitudinal segments parallel to the overall dike trend and cross segments at roughly right angles to them. Cross segments are abundant but invariably short, and most are thin; many are mere stringers.

3. Numerous cross segments terminate at one or both ends against longitudinal segments and in some places link those segments at fairly regular intervals of 20–50 cm. The converse relation, where longitudinal segments terminate against cross segments (points E, F) or show abrupt changes in thickness across them (point G), is also common. Nowhere did we observe longitudinal and cross segments cutting each other; their relation is consistently one of termination rather than intersection. Together these relations imply that the longitudinal and cross fractures developed at about the same time and thus are products of the same deformation. Had one fracture set predated the other, then the fractures of the later set would consistently terminate against, or cut across, those formed earlier.

4. Sinuous longitudinal segments tend to thin wherever they deviate markedly from the overall dike trend (points H, I, J), a geometry suggestive of separation of dike walls almost perpendicular to the main dike trend rather than at some angle oblique to it.

5. The several places marked by arrows where dike segments appear to end abruptly do not represent true terminations but rather are places where the top of a segment plunges below the trench floor (D, down) or the base of a segment emerges from it (U, up). Angles of plunge are highly variable, from gentle to steep (70° or more), the latter more common. Lateral extensions of several plunging dike segments were confirmed by digging and suggest that more interconnections exist among the various segments than is evident in figure 16.

6. Individual dike segments commonly split laterally into two or more segments, some of which converge again to enclose lensoidal masses of rock (points K, L, M).

7. Many small dike segments that split off from larger ones taper northwestward to zero thickness (left on figure), suggesting that the overall direction of dike propagation in this area was to the northwest. This is

Figure 18. Eroded cleft along gilsonite dike of Pariette system in mudstone of Uinta Formation; NE1/4 sec. 5, T. 9 S., R. 17 E., along low bluff bordering stream 60 m south of Duchesne County Road 216. Dike in mudstone consists of multiple thin segments connected at various points by small sills; aggregate thickness is about 0.7 m. The same dike in overlying massive sandstone formed a single tabular mass about 0.4 m thick (open cleft at top of photograph). Hammer for scale.

consistent with the location of the trench, only 0.5 km from the northwest end of the 6.4-km-long dike.

8. Small, irregular fragments of wallrock detached from dike walls and embedded in gilsonite are common.

9. Excavation of the trench floor to trace plunging dike segments revealed at point N a thin sill of undetermined lateral extent.

The common presence of sills and the extreme vertical variability in dike-sill geometry are shown in figure 17, a map of one wall of a trench dug perpendicular to the same dike described above, but 1 km to the southeast. Sills in this vertical cut, though invariably thin, served as feeders to those dike segments that bottom against them and provided flow paths between other dike segments to which they are linked. Terminations of sills against dike segments and of dike segments against sills are both common, but, as with the orthogonal network of dike segments shown in the previous figure, no intersections were found. These relations again suggest that all of the gilsonite-filled fractures—the longitudinal segments, cross segments, and sills—formed together during the same intrusive event. The curiously shaped mass at point A on the trench wall seems to represent the blunt, rounded end of a sill; it penetrates no farther than several inches into the rock. Its bulbous shape and the deformation of the rock around it show once more that the mudstone was not yet completely lithified at the time of intrusion.

Inspection of other trenches in the area and of additional dikes of the Pariette system (fig. 18) suggests that the relations documented above are typical of gilsonite dikes in mudstone of the Uinta Formation. Individual dike segments generally are thin (15 cm or less), sinuous, branching, and relatively short; only rarely can one segment be traced in continuity for more than 15 m laterally. Cross-dike segments and sills are both common but are thinner and shorter on average than longitudinal dike segments. These three structural elements form highly interconnected, orthogonal networks within tabular, vertical zones generally 2–3 m wide. The same dikes in sandstone, however, have a much different and more simple geometry; generally they show fewer branch- and cross-dike segments and are far longer, more continuous, and straighter than those in mudstone. The dike shown in figure 16, for example, has a completely different character where it enters a sandstone lens 10 m southeast of the part shown. At that point it becomes a single dike 36 cm thick; immediately beyond, it was sufficiently thick to mine and its former presence is marked by a rectilinear trench 37 m long. The sandstone wallrocks to either side reveal no trace of subsidiary dikes or cross segments. The walls of this and nearby dikes, except where modified by detachment of fragments and slabs of wallrock during intrusion, are smooth and almost planar, revealing no sign of plastic deformation. These relations again suggest to us that during intrusion of the gilsonite the sandstones were fully lithified but some of the mudstones were not.

Dike and sill geometries in older stratigraphic units, in contrast to those just discussed, show no sign that the fine-grained rocks were incompletely lithified during intrusion. Figures 19 and 20 show a sill complex associated with two small dikes near the base of the Rainbow dike, within the Green River Formation about 18 m above the top of the Mahogany ledge. Finely laminated marlstone, lean oil shale, and silty fine-grained sandstone are the dominant rock types at this outcrop. During intrusion the shale and marlstone split freely along the laminae so that the upper and lower contacts of each sill are smooth and parallel to bedding, except where they jump section to a higher position (fig. 20). Sheets of rock bounded above and below by sills were in places broken further by thin,

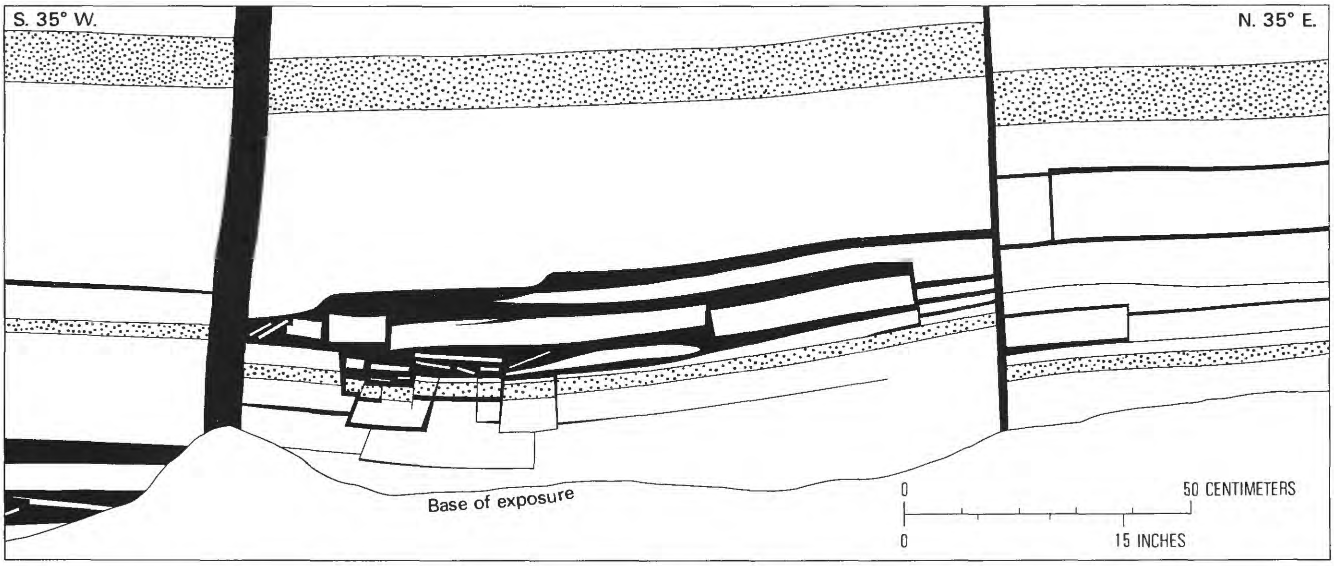

Figure 19. Vertical section of sill complex (solid) associated with two dikes (solid) near southeast terminus of Rainbow dike of Rainbow system, SW1/4 sec. 30, T. 11 S., R. 25 E. Field sketch of steep outcrop face in marlstone and lean oil shale (unpatterned) and sandstone (dotted) of upper part of the Green River Formation; view northwest. Figure 20 is photograph taken immediately south (left) of section shown. Modified from Verbeek and Grout (1992, fig. 9).

vertical dikelets; during continued intrusion and thickening of the sills the resultant blocks and slabs became embedded at various angles in the gilsonite. This dike-sill complex, though of comparable dimension to that shown in figure 17, nevertheless has fundamentally different geometry. Missing are the irregular, branching dike segments with their sinuous walls, the similarly irregular walls of the associated sills, the bulbous apophyses of gilsonite bordered by plastically deformed mudstone, and the anhedral clasts of wallrock embedded in the gilsonite. Instead, smooth-walled dikes and sills containing embedded blocks and slabs of rectangular form attest to the uniformly brittle response of the rocks to gilsonite intrusion in this area. Also shown in figure 19, and a fairly common feature of dike-sill complexes in the entire region, are minor stratigraphic mismatches across the dikes; these resulted from differential vertical uplift during intrusion of the underlying sills. Though not apparent at this locality due to the small interval exposed, in other areas the amount of offset can be shown to vary vertically as a function of the difference, from one side of the dike to the other, in the cumulative thickness of the sills injected at various stratigraphic levels beneath.

FAULTING OF DIKES

New advances in understanding the morphological features of extension fractures in rock and the effects of fracture propagation through heterogeneous sequences of strata have proven useful in reinterpreting some features of gilsonite dikes earlier attributed to postdike faulting. Eldridge (1906, p. 441), for example, reported that the walls of gilsonite dikes "frequently show slickensides indicating a movement in the direction of the markings," but he also noted, in this and earlier papers, that evidence of relative displacement of the strata on opposite sides of the dikes is rare. Though much of the apparent movement postulated by Eldridge was by strike slip and thus would have resulted in minimal offset of the almost horizontal strata, the reported slickensides are a probable mistake for plumose structure, about which little was then known. That Eldridge reported slickensides as common in 1906 but as absent in his earlier work of 10 years before perhaps is due to the fact that both plumose structure and slickenside striations are best visible at grazing sun angles and commonly are quite inconspicuous under other conditions.

Lateral offsets or jogs of dike walls similar to that shown in figure 6 have been noted by many workers and most commonly are ascribed to postdike movement along transverse faults (Eldridge, 1896a, b, 1901; Murray, 1949; Pruitt, 1961; Cashion, 1967). Reported amounts of offset are uniformly small, typically less than 1 m and everywhere less than the dike thickness. As noted previously, however, most such jogs result from the filling of parallel but noncoplanar fractures and are original features of fracture-zone geometry. Of those dike-wall offsets investigated by us, none is associated with a definable fault surface in wallrocks to either side of the dike, nor does the brittle gilsonite between show any obvious sign of disturbance. Similarly, the en echelon geometry of some dikes of the Ouray system and probably also that of the Pariette dike (Pruitt, 1961) is primary, much as in some igneous dikes. The Northeast Ship Rock dike at Ship Rock, New Mexico (Delaney and Pollard, 1981), and several dikes of

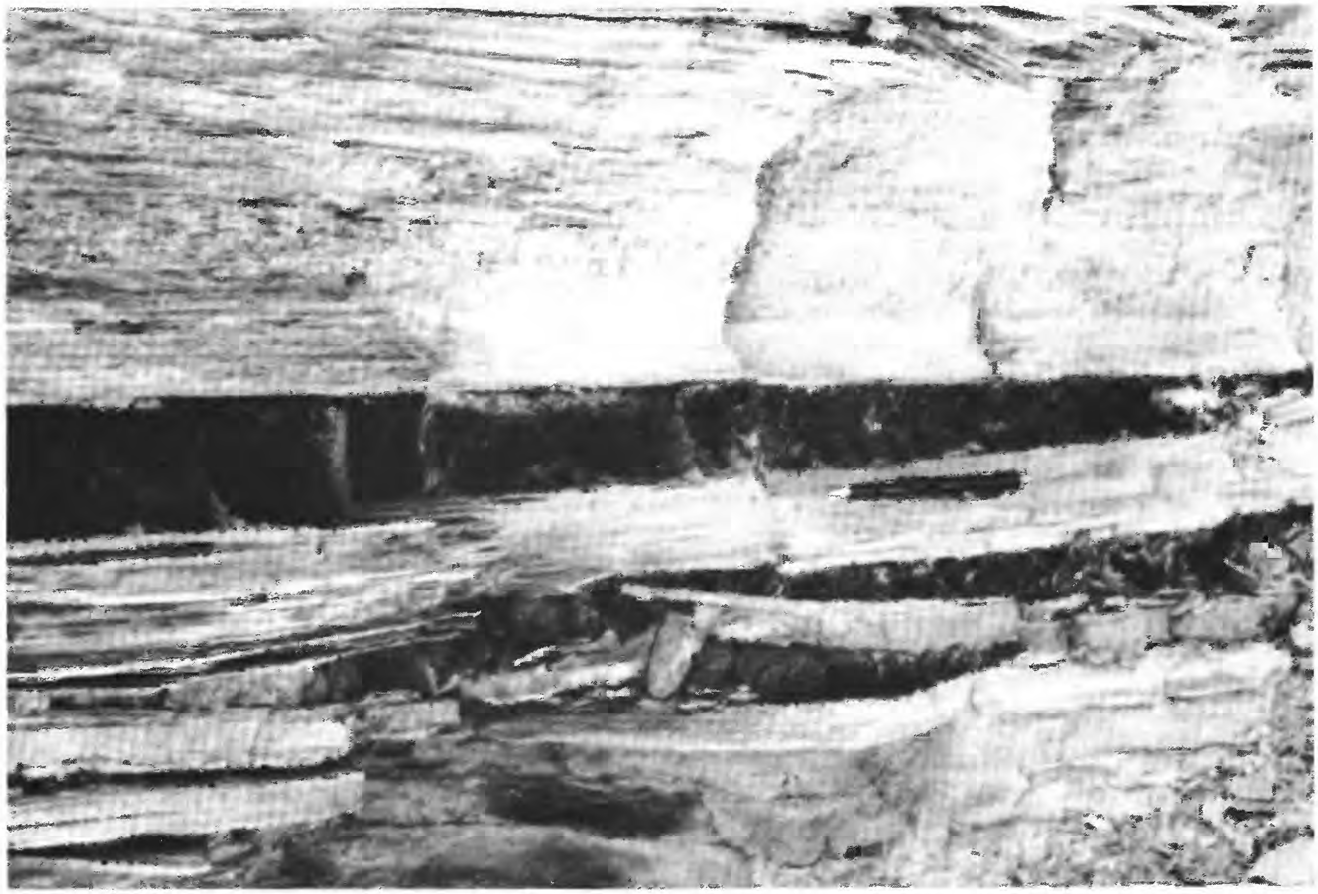

Figure 20. Gilsonite sills in finely laminated marlstone and lean oil shale of upper part of the Green River Formation, near southeast end of Rainbow dike of Rainbow dike system, SW¼ sec. 30, T. 11 S., R. 25 E. The sills 1–2 m farther northeast (right in photograph) connect to the dikes shown in figure 19. Note that some sills cut upsection and that slabs of oil shale are broken, slightly offset, and rotated within them. Pencil is 14 cm long.

the Spanish Peaks in southern Colorado (Smith, 1975) are familiar examples.

The splitting along bedding of dike fractures during propagation to produce almost horizontal steps on dike walls (fig. 12) is an additional feature sometimes mistakenly attributed to faulting. Early mention of these features was made in the careful and comprehensive reports of Eldridge (1896a) and Douglass (1928). In reference to one such supposed bed-parallel offset of the Black Dragon dike, Eldridge (1896a, p. 937) admitted that there was "little evidence of actual crushing" of the gilsonite along the trace of the inferred fault. Those offsets inspected by us likewise show no evidence of faulting of the gilsonite or of shear along bedding in the vicinity of the wallrock step.

Finally, minor stratal offsets across dikes are common in some areas, but evidence of faulting to produce them rarely is evident. Douglass (1928, p. 102), for example, noted of one locality along the Black Dragon dike that strata to one side were offset 3.5 m with respect to strata on the other but that the observed offset seemed not to persist downward, perhaps because the movement was "compensated by the soft shales." More likely, however, the vertical variation in amount of stratal mismatch is due to multiple injected sills extending laterally from the dike, as described in the previous section for the Rainbow dike. The area Douglass described is near the southeastern end of the Black Dragon dike, probably in the SW¼SW¼SW¼ sec. 13, T. 12 S., R. 25 E., about 1 km southeast of Dragon Canyon, where sills are particularly common within the basal beds of the Green River Formation. Other reports of vertical offsets along gilsonite dikes (Murray, 1949) probably are also to be ascribed to injection of sills; nowhere have we observed such an offset due to postdike faulting.

Evidence of faulting along gilsonite dikes was found by us at only a single locality, along County Road 207 about 1.6 km due north of the abandoned mining camp of Rainbow, in NE¼NE¼SW¼SW¼ sec. 13, T. 11 S., R. 24 E. Here, in a small roadcut, several thin, vertical gilsonite dikes contain irregular fractures lined with colorless to white calcite. Both the gilsonite and later calcite are scored by slickenside striations plunging between 0° and 19° SE., indicating nearly strike-slip movement.

MINERALIZATION AND ALTERATION OF DIKE WALLS

Alteration of the wallrock adjacent to gilsonite dikes and mineralization of the dike walls suggest that aqueous fluids, rather than free hydrocarbons, were the first fluids to migrate through the fractures now occupied by gilsonite.

Deposition of Limonite

Dike walls in all areas examined, both at the surface and deep underground in active mines, are discolored by limonite that precipitated within the pores of the wallrock. The discoloration in some places is so pronounced that the wallrock is dark ochre brown to almost black, yet nearby joints of regional sets in the same rocks typically are little stained, if at all. Though much of the limonite was precipitated in available pore space in the wallrock bordering the dikes, additional limonite on some dike walls forms a discrete coating between the wallrock and the attached gilsonite. The limonite is thus the earlier deposit and not a late precipitate (Monson and Parnell, 1992; Verbeek and Grout, 1992).

The early age of the limonite coatings is perhaps best appreciated where fragments of detached wallrock are embedded in gilsonite. Near the northern end of the Black Dragon dike, for example, diversely oriented clasts of marlstone within the gilsonite (fig. 15) are most prominently coated with limonite on those surfaces that formerly were parallel to the dike walls and locally to bedding, but surfaces oblique to both show no such coatings. Such preferential staining of wallrock clasts embedded in almost impermeable gilsonite argues strongly for a pre-gilsonite age of much of the limonite at this locality. Similar relations were documented for a dike of the Pariette system at the outcrop shown in figure 18.

Deposition of Calcite

Calcite deposited on the walls of gilsonite dikes was documented by John C. Lorenz (written commun., 1992) for several dikes in the western part of the dike province. Photographs supplied by him show remnant patches of gilsonite adhering to sheets of white calcite 1–2 mm thick, the latter in turn still firmly attached to the dike walls. In most places the gilsonite intruded the mineralized fractures along the median suture of the calcite vein, but in one place the gilsonite broke across the calcite and intruded along the vein wall, leaving the entire thickness of the calcite fill attached to the opposing wall. The gilsonite is the later deposit. Similar observations and conclusions were made by Monson and Parnell (1992).

Bleaching of Dike Walls

Bleached rock bordering gilsonite dikes was seen in most areas examined and is the most obvious manifestation of wallrock alteration. The bleached zones range in thickness from a few millimeters in fine-grained beds to as much as 50 cm in some sandstones. Brief descriptions of their appearance in several areas follow.

1. Mudstone in the area of the dike shown in figure 16 is of somewhat variable color, medium grayish brown for the most part but locally reddish brown to medium brownish maroon. Regardless of original color, the wallrock bordering the gilsonite dikes has been bleached and impregnated with limonite for distances of 0.5–1.5 cm from the dikes. The bleached zones are especially conspicuous in the reddish-brown to maroon mudstone, which adjacent to the dikes has lost all vestige of red color.
2. Altered zones 2–4 cm wide border the dikes and sills shown in figure 17; within these zones the normally reddish-brown mudstone is prominently bleached to pale olive and heavily stained with limonite. A full return to normal color is not evident until about 4–6 cm from the larger dikes and about half that distance from the sills.
3. A thin (8 cm) bed of purplish-gray sandstone next to a mined dike in Cottonwood Wash (Pariette system) has been bleached for a distance of 47 cm from the dike walls.

Though no geochemical study has yet been made of these altered zones, the evident destruction or removal of original hematite pigment and the widespread deposition of hydrated iron oxides are obvious visual indicators of the redistribution of iron by aqueous fluids circulating through the dike-fracture network. Continuity of the bleached zones along fractures discontinuously occupied by gilsonite further suggests that the bleaching, similar to the deposition of the associated limonite and calcite, occurred before intrusion of the gilsonite into the fractures. Again the dikes of the Ouray system serve as convenient examples. Three sills and a dike segment of the complex shown in figure 17 pinch out completely within the mudstone, but continuing beyond each of them is a closed fracture (dashed lines on figure) devoid of gilsonite but bordered by bleached and limonite-stained rock. Widths of the altered zones show no relation to the thickness, or even to the presence, of the gilsonite but are effectively constant for each fracture along its length. Analogous features were noted (but not specifically mapped) among the dike segments of figure 16 and along other dikes not shown here, and in many places fractures within a meter or two of individual dike segments

Figure 21. Sandstone wallrock impregnated with gilsonite adjacent to thin dike, SW1/4 sec. 13, T. 12 S., R. 25 E. Southeast end of Black Dragon dike (Rainbow system), near its base in clastic rocks of lower part of the Green River Formation.

resemble those segments in every respect but for the absence of gilsonite. From these observations we conclude that gilsonite dikes resulted from the incomplete intrusion of narrow zones of pre-existing fractures that formerly had channeled aqueous fluids through a considerable thickness of the sedimentary section. The genesis of these zones as hydraulic fractures due to fluid overpressures in the hydrocarbon source beds is discussed in a later section.

Deposition of Chlorite in Sandstone Wallrock

The diagenesis of sandstone wallrock adjacent to gilsonite dikes was studied in 36 samples by Monson and Parnell (1992) using thin-section and scanning electron microscopy techniques. The diagenetic sequence of these rocks is similar to that reported previously by Pitman and others (1982) for sandstone reservoir rocks on a regional basis except for the composition of the authigenic clay component, which normally is illite and mixed-layer clays but near the dikes is mostly chlorite. Much of the chlorite grew on detrital sand grains and within pore throats as well-developed rosettes of bladed crystals; the crystals subsequently were enveloped by gilsonite, which filled all remaining pore space. Monson and Parnell (1992) interpreted their findings as evidence for the passage of magnesium-rich fluids through the dike fractures before the same fractures were invaded by gilsonite. Authigenic quartz crystals nucleated on individual wallrock clasts embedded in gilsonite were noted by the same authors and provide additional evidence for the early passage of aqueous fluids through the dike-fracture system.

Impregnation of Dike Walls by Gilsonite

Gilsonite during intrusion showed only a limited tendency to saturate adjacent wallrocks. A sharp contact between pure gilsonite and dike walls is common wherever the wallrocks are relatively fine grained and pore sizes correspondingly small; observed widths of zones of gilsonite impregnation in such rocks range from only a fraction of an inch (Davis, 1957) to nil. In sandstone, however, saturation of the wallrock by gilsonite is more common, an effect often noted before (Eldridge, 1896a; Murray, 1949; Davis, 1957; Pruitt, 1961; Crawford and Pruitt, 1963; Cashion, 1967) and illustrated in figure 21. Reported distances of impregnation generally are 1 m or less, commonly much less, but are as much as 6 m in places along the Pariette dike (Pruitt, 1961). The minor extent to which gilsonite was able to invade most strata was attributed by Monson and Parnell (1992, p. 262) to prior precipitation of chlorite cement, the abundance of which near dike walls "severely restricted the permeability of the host sandstones."

Within the bird's-nest zone of the upper part of the Green River Formation, rounded, oblate cavities generally 5–20 cm across are filled with gilsonite adjacent to the Cowboy (Eureka) dike. Similar cavities nearby, in the same beds but not in contact with the dike, are filled instead with calcite. The cavities resulted from dissolution of globular aggregates of a pre-existing authigenic mineral, probably nahcolite (Milton and Eugster, 1959); features of identical appearance but still containing nahcolite are common in stratigraphically equivalent beds in some of the underground oil-shale mines farther east, in Colorado. Field relations along the Cowboy dike show that the nahcolite dissolved before the dike was filled but do not indicate whether dissolution was a local phenomenon (due to early aqueous fluids circulating through the dike-fracture network) or a more general effect of regional groundwater circulation.

Table 1. Strike, relative abundance, and stratigraphic distribution of regional joint sets in the eastern Uinta basin, Utah and Colorado

Set	Strike range	Abundance	Stratigraphic distribution (formation)
F_1 (oldest)	N. 15°–30° W.	Sparse[1]	Wasatch[2]
F_2	N. 55°–85° W.	Very abundant	Duchesne River
		Very abundant	Uinta
		Very abundant	Green River
		Moderate	Wasatch
F_3	N. 60°–80° E.	Moderate	Duchesne River
		Moderate	Uinta
		Moderate	Green River
		Moderate	Wasatch
F_4	N. 15°–40° E.	Sparse	Duchesne River
		Very abundant	Uinta
		Very abundant	Green River
		Sparse	Wasatch
F_5 (youngest)	N. 65°–85° W.	Sparse	Duchesne River
		Sparse	Uinta
		Sparse	Green River
		Sparse	Wasatch

[1]F_1 set increases in prominence westward and is dominant set in much of region west of the gilsonite dikes.

[2]Present also in Green River, Uinta, and Duchesne River Formations farther west.

RELATION OF DIKES TO FRACTURE NETWORK OF HOST ROCKS

Regional Fracture Network of Eastern Uinta Basin

Joint networks in the Wasatch, Green River, Uinta, and Duchesne River Formations were documented at 119 sites in the Uinta basin (70 within the study area) during the course of this study to determine the relation, if any, between regional fracture sets and gilsonite dikes in the same rocks. Field methods used to identify and characterize the joint sets and to determine their relative ages are described in Grout and Verbeek (1983) and Grout (1988). Five sets of vertical extension joints were identified in the eastern Uinta basin, although at most outcrops only two sets, and locally three, are present. These five sets are elements of a regional fracture network that extends well into western Colorado and that covers a minimum known area of more than 30,000 km^2. A brief overview of that network is given here to place the gilsonite dikes within a broader context.

For convenience, the five regional joint sets are referred to, from oldest to youngest, as F_1 through F_5. Their orientations, relative abundance, and stratigraphic distribution in the eastern Uinta basin are given in table 1. Each set can be matched to its probable counterpart in correlative rocks farther east, in the Piceance basin, on the basis of orientation, sequence of formation, and style; the fracture histories of the two areas are comparable, and the same symbology of fracture sets is used for both. Much of the recently published information on fractures in the Piceance basin (Grout and Verbeek, 1983, 1985, 1992; Verbeek and Grout, 1983, 1984, 1986) applies as well to the area of the gilsonite dikes.

Joints of the oldest, F_1 set strike north-northwest to north and within parts of the Uinta basin are the dominant set, as in much of the area between Duchesne and Price (fig. 1), west of the area occupied by the gilsonite dikes. Within the eastern Uinta basin, however, they wane in prominence and have been found to date only in strata of the Wasatch Formation near and along the Utah-Colorado State line, east of and stratigraphically below all but a few of the gilsonite dikes. In the Piceance basin, too, they are widely scattered but generally subordinate to other sets. They will be considered no further here other than to note that, after F_1 time, the host rocks for the gilsonite dikes still were unfractured within the region occupied by the dikes. The later F_2 set, in contrast, affected the whole of the Piceance and eastern Uinta basins. These northwest- to west-northwest-striking joints are almost ubiquitous and were the first to form in most places; on a regional scale they are products of the strongest period of post-Laramide extensional deformation to have affected the Cenozoic rocks. Their probable age based on stratigraphic and geomorphic evidence is somewhat loosely bracketed as 36–10 Ma (Grout and Verbeek, 1992; this paper). The prominence of the F_2 set gradually wanes southward, however, and the set is only weakly expressed over large parts of the southern Piceance basin. Joints of the next younger, F_3 set strike east-northeast and tend to have formed in greatest abundance wherever the F_2 joints were least developed; thus the F_3 joints are the dominant set in the southern Piceance basin (Grout and Verbeek, 1985; Grout, 1991), are almost absent from the northern part of the same basin (Verbeek and Grout, 1983, 1984), and are widespread but generally subordinate to the F_2 set in the eastern part of the Uinta basin. The F_2 and F_3 sets together are the dominant components of the regional fracture network; the gradual regional shifts in their relative prominence reflect gradually changing stress gradients through time (Verbeek and Grout, 1986).

Younger sets seem to be products of regional uplift, the main phase of which in the Piceance basin began 8–10 Ma (Whitney and Andrews, 1983). Joints of the F_4 set are particularly widespread, though in most areas subordinate in size and commonly in abundance to members of earlier sets. The F_4 joints strike north-northeast through north to north-northwest in different places, depending mostly on

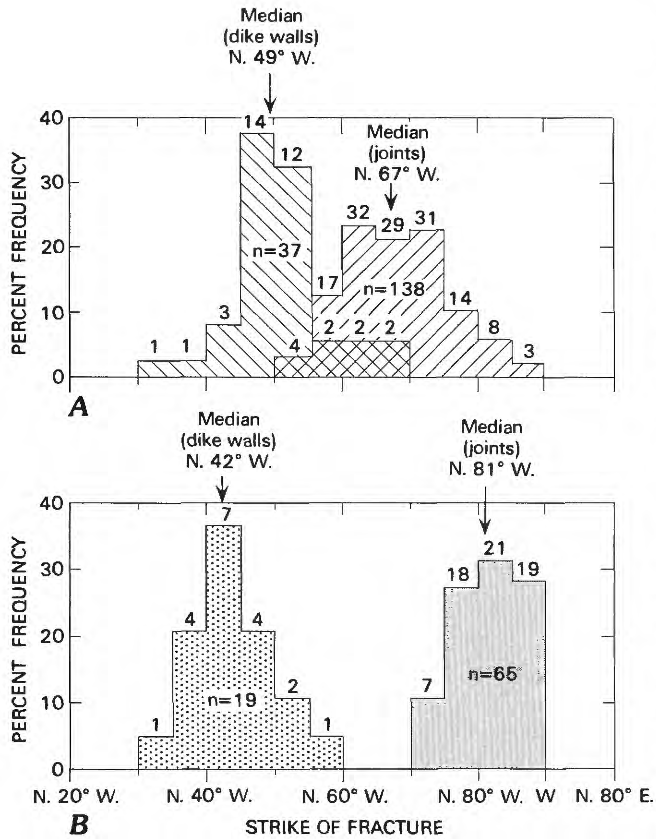

Figure 22. Histograms of orientations of dike walls and of F_2 joints in nearby outcrops for Ouray (*A*) and Pariette (*B*) dike systems. Note angular discordance between dikes and joints. Numbers above histogram bars represent number of readings within each 5° strike interval. Modified from Verbeek and Grout (1992, fig. 14).

which sets were already present (Verbeek and Grout, 1984). Closely associated with the F_4 joints are almost horizontal, bed-parallel fractures that formed during erosional unloading of the strata; these are the sedimentary equivalents of sheeting joints in massive plutonic rocks. Both types of joint are common in the eastern Uinta basin. The youngest tectonic joints to form, members of the F_5 set, are widely scattered both in the Piceance and Uinta basins but only rarely are prominent elements of the local fracture network; they are parallel to joints of the F_2 set but are much smaller and younger, and formed at shallower depths (Verbeek and Grout, 1983).

Age of Dikes Relative to F_2 Joint Set

Three joint sets (F_2, F_3, F_4) among the five identified in the study area are of wide distribution. Our attention here is directed mostly to the F_2 set because its joints in many places dominate the fracture network and strike almost parallel to the gilsonite dikes (plate 1*B*), inviting the hypothesis that the dikes originated through invasion of previously formed joints (see Davis, 1957, and Cashion, 1967, for a summary). Three lines of evidence, however, collectively show that the fractures occupied by the gilsonite dikes are not part of the regional joint network and, furthermore, that the regional joint sets postdate the dikes. The evidence includes differences in (1) orientation, (2) physical characteristics, and (3) abutting relations between dike walls and joint surfaces.

The issue of parallelism is the most easily dealt with because the results of our measurements reveal a close, but nevertheless inexact, correspondence in orientation between dike walls and joints in the adjacent host rocks. The degree of angular discordance generally is on the order of 10° or less, too small to be readily apparent except under conditions of favorable exposure, but locally it is 20° or more and is then obvious in the field. Data from two such areas are summarized in figure 22, which shows orientations of dike walls for dikes of the Ouray (*A*) and Pariette (*B*) systems as compared to F_2 joint orientations in the local host rocks. To facilitate comparison, only data from sandstone beds were used because it is in these rocks that the dikes are most nearly planar and the joints best developed. The distributions for the Ouray examples show only minimal overlap, and none within one standard deviation of their respective means: fully two-thirds of the dike walls strike N. 45°–55° W., whereas two-thirds of the F_2 joints in the same rocks strike N. 60°–75° W. That the data appear to define two independent distributions whose means are separated by 18° suggests to us that dilation of pre-existing joints was not an important mechanism in the formation of the gilsonite dikes of this area. This conclusion is confirmed by data from the Pariette system (fig. 22*B*), for which the distributions show no overlap and the median orientations of dike walls and nearby F_2 joints are separated by 39°. Similar mismatches in orientation between dike walls and nearby F_2 joints were observed along some of the dikes of the Rainbow system.

A second argument is that dike walls and joints in many places are dimensionally and morphologically different and their spatial characteristics dissimilar, even where the two share a common orientation. Individual dike segments at almost all localities visited can be traced in lateral continuity for distances significantly greater than even the longest joints in the adjacent host rocks. Dike segments more than 5 m long are common throughout the region, even in mudstone where multiple splits and abrupt terminations during fracture propagation severely limited the length to which an individual dike segment could grow; two of the dike segments shown in figure 16, for example, have exposed lengths of 6–9 m. Lengths of 10–20 m are common in nearby sandstones, and even longer individual dike segments are hardly rare throughout the region. A review of our data for 119 joint stations, however, shows that regional F_2 joints parallel or subparallel to the dikes, though abundant, generally are only

Figure 23. Joints of F_2 fracture set in thin sandstone bed of Uinta Formation, about 6.5 km south-southwest of the town of Ouray, SW1/4 sec. 17, T. 9 S., R. 20 E. Joints strike at low acute angle to nearby gilsonite dike (off photograph to right). Note short length of joints; hammer for scale.

2–4 m long and frequently shorter (fig. 23). It is evident that intrusion of gilsonite into existing joints would have produced much shorter individual dike segments than those observed. Lengthening of pre-existing joints during intrusion is an unlikely explanation for the dimensional discrepancies because the morphologies of individual dike segments and their spacings and interconnections in many areas are incompatible with derivation from an initial set of joints. The large twist-hackle faces on the walls of the Cowboy dike (fig. 11), for example, have no size counterpart among the regional F_2 joints; the joints and the dike walls are wholly unlike in overall style. Similarly, the highly irregular and branching dike segments common among dikes in mudstone of the Uinta Formation (fig. 16) show no evidence of having originated from, or been influenced by, the almost planar to gently sinuous regional joints in the same rocks. At many localities one is hard pressed to imagine how the dikes could have formed from the existing fracture network and is led instead to conclude that the dikes formed first and the joints later.

Abutting relations between longitudinal- and cross-dike segments further emphasize the separate genesis of the gilsonite-filled fractures versus the regional joint sets. Key properties of the gilsonite-filled fractures (fig. 24*A*) are that (1) terminations of cross-dike segments against longitudinal dike segments are common; (2) terminations of longitudinal segments against cross segments are also common; and (3) no intersections, only terminations, were observed. Such a geometry implies either coeval formation of the two sets of segments (the interpretation favored here) or a three-stage process involving the formation of each set in turn, followed by reactivation (renewed growth) of the earlier set to form fracture terminations in the opposing senses. Regional joint sets in the adjacent rocks, however, have the geometry shown in figure 24*B*: cross joints (F_4) either terminate against or cut across longitudinal joints (F_2), but the converse was not observed because the fractures of the one set uniformly postdate those of the other. The dike and joint geometries, though superficially similar in some places, nonetheless are fundamentally different. Key differences between them—in particular, the abundance of crosscutting relations among the joints versus their absence among the dike segments—suggest that the one could not have been derived from the other.

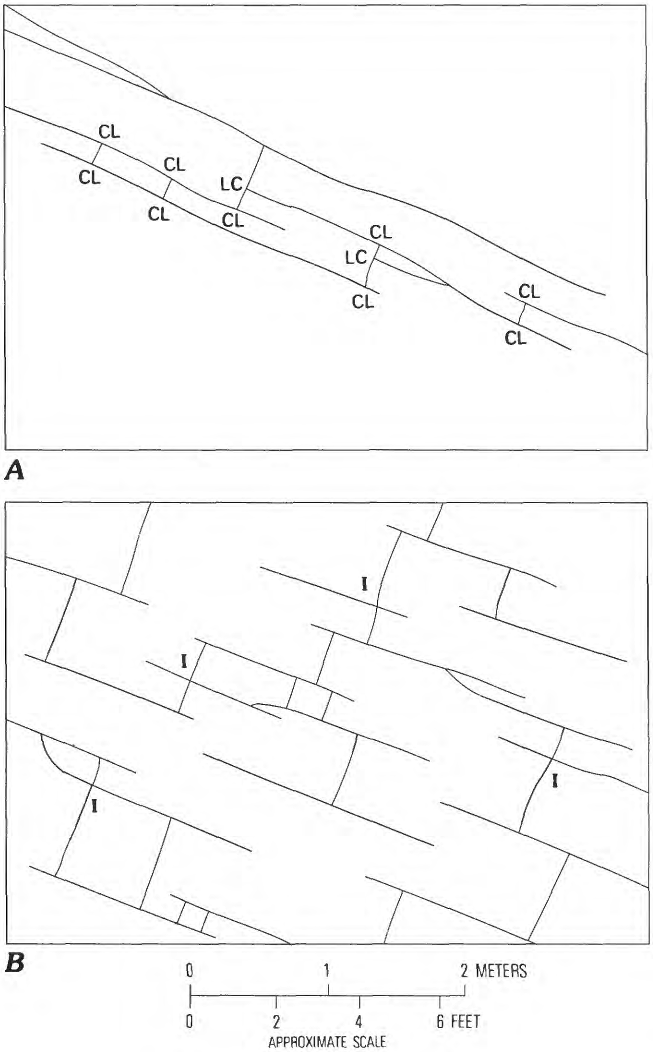

Figure 24. Sketch showing difference in fracture pattern between (*A*) longitudinal and cross-dike segments and (*B*) F_2 and F_4 joints. CL indicates terminations of cross-dike segments against longitudinal segments, LC signifies the converse, and I indicates fracture intersections (crosscutting fractures). Younger fractures that cut across older, mineralized fractures are common among joints but are absent from dikes. Modified from Verbeek and Grout (1992, fig. 15).

In summary, the gilsonite-filled fractures differ from those of regional joint sets in size, in overall shape, in abutting and crosscutting relations, and locally in orientation. We find no evidence that dike geometries were in any way influenced or controlled by pre-existing joint sets and conclude that those sets were absent at the time of intrusion.

Pencillated Structure in Dikes

Gilsonite pervaded by small fractures and that breaks into rude columns (fig. 25) or plates (fig. 26) instead of the usual conchoidal fragments has long been described as "pencillated." The term was attributed by Wurtz (1869) to one Professor Lesley, who about 1864 used it in reference to the columnar fracture exhibited by a dike of grahamite, an asphaltite related to gilsonite, in West Virginia. Pencillated structure in thin (<20 cm) gilsonite dikes commonly extends from one wall to the other, but in very thick dikes it may penetrate no farther into the gilsonite than 15 cm or so (Davis, 1957; Cashion, 1967). Hypotheses on the origin of pencillated structure in gilsonite are diverse and include cooling (Eldridge, 1901), dehydration or loss of volatiles (Douglass, 1928; Crawford, 1949), near-surface weathering (Davis, 1957; Pruitt, 1961), multiple episodes of intrusion (Davis, 1957), and compression due to movement of the enclosing rocks (Eldridge, 1902; Douglass, 1928; Cashion, 1967). Many of the fractures we have observed in gilsonite, however, are members of the same regional joint sets that formed in the host rocks. Most common within dikes are members of the F_3 and F_4 joint sets (fig. 27), which divide the gilsonite into thin (3–10 mm) vertical slabs (fig. 26) oblique or perpendicular to the dike walls, respectively. The slabs commonly are

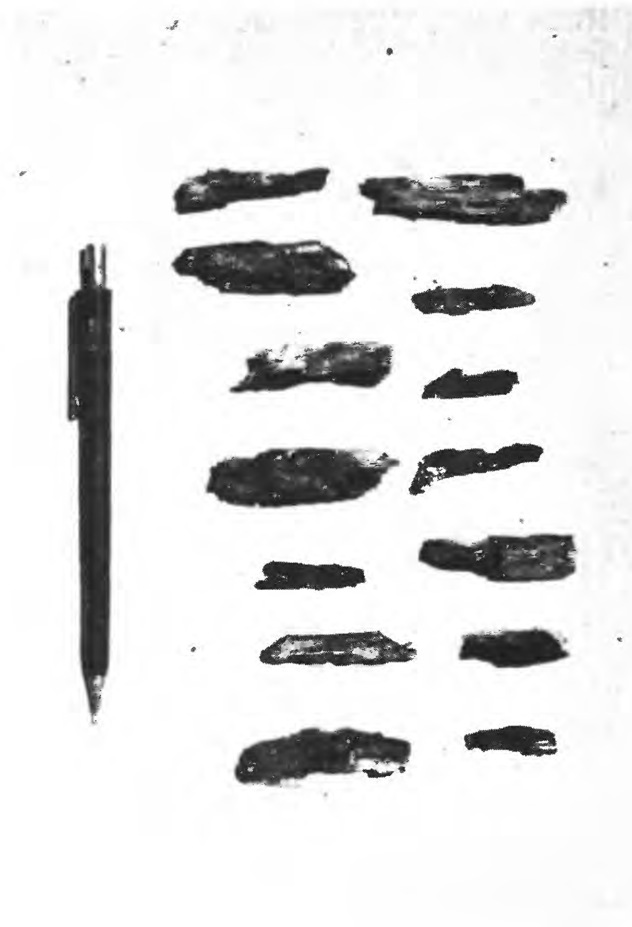

Figure 25. Gilsonite pencils extracted from dike of Ouray system in SW¼ sec. 17, T. 9 S., R. 20 E. Mechanical pencil at left is 14 cm long.

Figure 26. Platy structure in small dike of Pariette system near Castle Peak, NW1/4 sec. 4, T. 9 S., R. 17 E. Plan view, north-northwest toward top of photograph. Fractures bounding plates strike N. 33° E. and correspond to F_4 joint set in adjacent clastic rocks.

broken further by gently dipping fractures equivalent to the bed-parallel unloading joints in adjacent strata. The subhorizontal columns so defined (fig. 25) probably correspond, in part, to the pencillated structure of earlier writers.

East-northeast-striking joints of the F_3 set are particularly widespread (fig. 27) within gilsonite dikes, though their prominence varies widely and some dikes lack them completely. The appearance of ore from the Harrison vein of the Rainbow system, described by Douglass (1928, p. 88) and Pruitt (1961, p. 39) as splitting "into plates diagonal to the vein," possibly is to be ascribed to these joints. Where present the F_3 joints typically strike N. 55°–85° E. and are spaced 3–10 mm apart in the brittle gilsonite; their counterparts in the enclosing host rocks are much more widely spaced at 0.5–3.0 m. In some places the F_3 joints in the gilsonite are curved, a probable result of stress reorientation near the dike walls: joints in the dike interior strike parallel to the local average for the F_3 set, whereas those nearer the dike margins curve to meet the dike walls at a high angle. A similar but more subtle effect of stress reorientation is a gradual swing in joint attitude as dikes narrow. Near the northwestern end of the Cowboy dike, for example, where the dike tapers in thickness from 20 cm to 4 cm in a distance of only 2 m, the F_3 joints gradually change strike from N. 55° E. to N. 15° E., the latter almost perpendicular to the dike walls and 40° removed from the average local strike of the F_3 set.

To what extent processes other than regional extension fracturing of gilsonite contributed to pencillated and kindred structures remains uncertain, and we have made no special study of them. Nevertheless, commonly used adjectives such as platy, cuboidal, and columnar reinforce our belief that systematic (nonrandom) regional fracturing of gilsonite is the dominant process.

Local Dike-Parallel Joints

Fractures parallel to gilsonite dikes, but not themselves filled with that substance, were observed along almost all dikes examined (Verbeek and Grout, 1992). These fractures, although similar in strike to joints of the F_2 set, especially in the eastern part of the study area (compare plates 1*B* and *C*), are not members of a regional joint set but instead are restricted, in most places, to narrow zones of wallrock bordering each dike (fig. 28). Abundances of 3–6 dike-parallel fractures per meter are common near dike walls in sandstone, but their numbers typically decrease almost to zero within 5–10 m from the dikes, as noted also by Monson and Parnell (1992). Plumose structure and local twist hackle and arrest lines show that the dike-parallel fractures are extension joints. Their surfaces, like those of the dike walls, commonly are coated with limonite, heavily so in some places, and the adjacent wallrocks are bleached. In this, as well as in their strict parallelism and close spatial relation to the dikes, they differ from the associated F_2 joints, with which they might otherwise be confused. The same properties strongly suggest that the dike-parallel joints are genetically linked to the dikes themselves.

Opportunity to determine the relative age of the local dike-parallel and regional F_2 joints is limited in most places by their similar orientations; abutting and intersecting relations are few. West of the Green River, however, where dikes of the Pariette system (N. 34°–55° W.) and joints of the F_2 set (N. 77°–84° W.) differ in strike by 30°–40°, numerous terminations of F_2 joints against dike-parallel joints (fig. 29) consistently establish the latter as the older fractures. Inasmuch as the still-older F_1 fracture set did not form in most of the region occupied by the gilsonite dikes, we conclude from these observations that the dike-parallel fractures were generated in unbroken rock. Their pre-F_2 age reinforces our belief that they formed during the intrusion process.

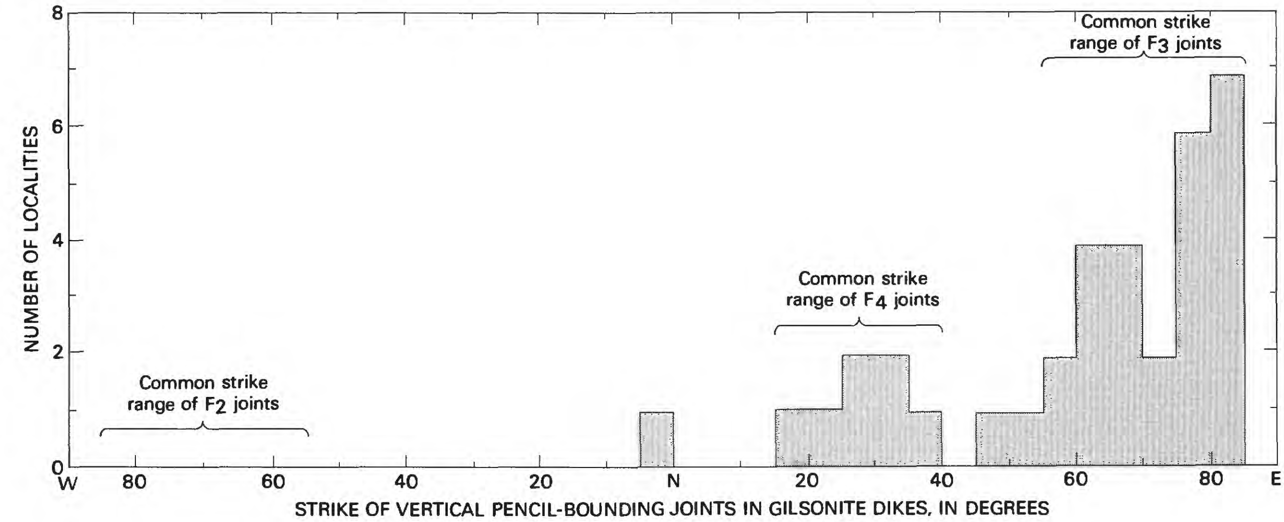

Figure 27. Histogram of orientations of vertical pencil-bounding joints. Pencils result from formation of two sets of closely spaced fractures in the brittle gilsonite. Fractures of one set generally are vertical and correlate with joints of either the regional F_3 or F_4 sets; fractures of other set are subhorizontal and formed upon regional unloading. Pencil axes thus commonly are horizontal and trend in direction of dominant vertical set. Note absence of F_2 joints as component of pencillated structure.

Fracture zones geometrically analogous to those bordering the gilsonite dikes have been documented elsewhere, most notably in association with mafic igneous dikes in the Four Corners region of New Mexico and in central Utah (Delaney and others, 1986). Similar fractures also are present in western Colorado, adjacent to the clastic dikes described briefly by Grout and Verbeek (1982); in Greenland, in association with mafic igneous dikes (Rogers and Bird, 1987); and in southern Israel, adjacent to dikes of dolerite (Baer and Beyth, 1990) and to dikes of basalt and trachyte (Baer and Reches, 1991). Despite the diversity in dike-filling materials in the various areas (mafic through intermediate magmas, viscous hydrocarbons, tuffaceous sediments derived from overpressured ash-fall layers), the dike-parallel joints typically are best developed and most abundant near the dikes and display increasing spacing away from them. A further shared characteristic is the small size of the dike-parallel fractures relative to the dike walls. Lengths of dike-parallel joints flanking the gilsonite dikes, for example, typically are not much different from those of regional joints in the same rocks. Delaney and others (1986) also remarked on the small size of dike-adjacent joints relative to the dikes themselves in the areas they examined.

Cautious interpretation of dike-emplacement mechanisms was advised by Delaney and others (1986, p. 4933) for areas where the regional fracture network includes a set parallel to the dikes because of potential difficulty in determining whether the dike fluids ascended along pre-existing joints or intruded self-generated fractures. In the eastern Uinta basin, however, the demonstrated different age and slight to moderate angular discordance between the dike-parallel and F_2 joints preclude the interpretation that the gilsonitic fluids merely invaded pre-existing members of a regional joint set where these were most closely spaced. We conclude instead, as did Delaney and others (1986), Rogers and Bird (1987), Baer and Beyth (1990), and Baer and Reches (1991) for some of the igneous dikes investigated by them, that the dike-parallel joints are products of the intrusion process. Analysis by Delaney and his colleagues of the mechanical and hydraulically generated stresses ahead of a propagating dike shows, for reasonable dike geometries and fluid pressures, that induced tensile stresses near dike tips are of sufficient magnitude to fracture the host rock. Moreover, the magnitude of the induced stress is greatest to either side of the dike plane rather than directly ahead of it, thereby promoting fracture of the adjacent wallrock. The analysis shows as well that in rock about to be intruded (1) the magnitude of the induced tensile stress increases as the dike tip approaches and (2) the position of maximum induced stress migrates closer to the plane of the dike, thereby accounting not only for the induced fractures ahead of the dike but also for their increasing abundance closer to its walls. The advancing dike continuously bisects the zone of fractures formed ahead of it to produce the final observed geometry: a thin, tabular zone of broken rock, the length and height of the zone much greater than its thickness, occupied by a dike along its midplane. Induced fractures close to the advancing dike may be invaded by the dike fluid,

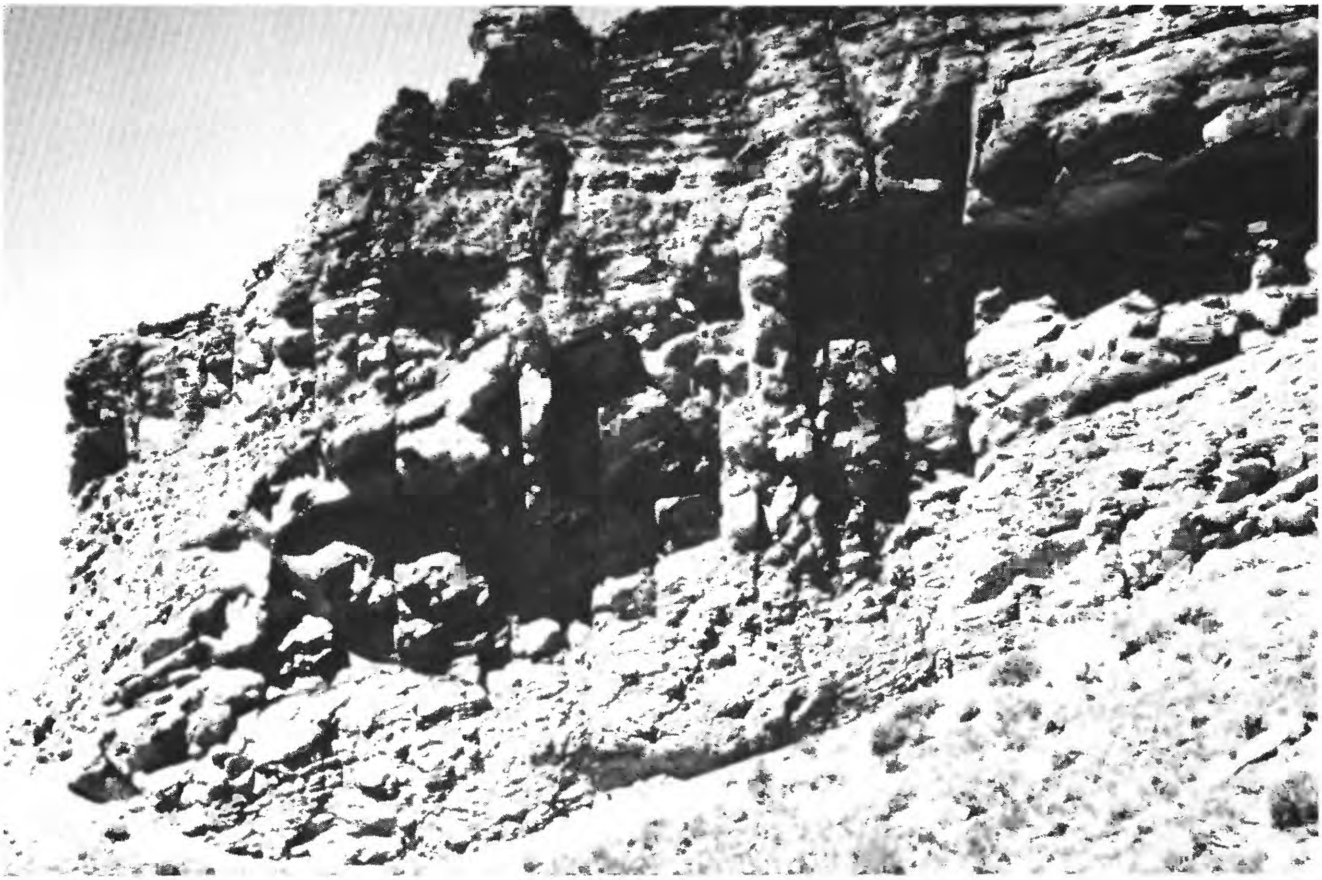

Figure 28. Zone of thin gilsonite dikes and dike-parallel joints cutting marlstone of Green River Formation and overlying, thinly bedded sandstone of Uinta Formation, NE1/4 sec. 5, T. 12 S., R. 24 E., about 6.5 km southwest of abandoned mining camp of Rainbow. Bluff is about 10 m high; view to northwest. Dip of beds is due to rapid loading of overlying sandstone during deposi-

but most others are not because they are not physically connected to the parent crack.

A complication to the geometry described above, but one valuable to understanding the regional geologic history, is that dike-parallel joints in some few strata are not confined to narrow zones flanking the gilsonite dikes but are present also at substantial distances away from them. At station 47 (plate 1*A*), for example, a single, thin (3–4 cm), well-cemented limestone bed within oil shale of the Green River Formation contains a set of closely spaced extension joints striking N. 40°–45° W., parallel to nearby gilsonite dikes of the Rainbow system but markedly different in orientation from the F_2 joints measured in numerous other strata (plate 1*B*). The limestone bed crops out at a point 105 m removed from the nearest known gilsonite dike, too distant for its joints plausibly to be ascribed to induced stresses during intrusion. In addition, near Castle Peak, where dikes of the Pariette system strike N. 34°–55° W., we found several beds that contain joints striking N. 40°–55° W. at distances as much as 2 km from the dikes. Joints of the F_2 set are present in some of these same beds and in most places crosscut the calcite-filled dike-parallel joints, but abutting relations at two localities confirm that F_2 is the younger set. Without exception, strata containing dike-parallel joints distant from the dikes are thin, well indurated, and fine grained; they are among the most brittle beds exposed. From this we infer (1) that the region was under northeast-southwest tectonic extension when the zones of gilsonite-dike fractures were forming, (2) that this regional stress was responsible for the strong northwest-preferred orientation of the dikes, and (3) that stress magnitudes beyond the local zones of stress amplification near the propagating fracture zones remained too low to fracture any but the most brittle rocks of the stratigraphic succession.

DEPTH OF EMPLACEMENT

Cashion (1967) noted that many dikes of the Cowboy-Bonanza and Rainbow systems terminate downward within or near the Mahogany zone of the upper part of the Green River Formation. Assuming that this relation holds

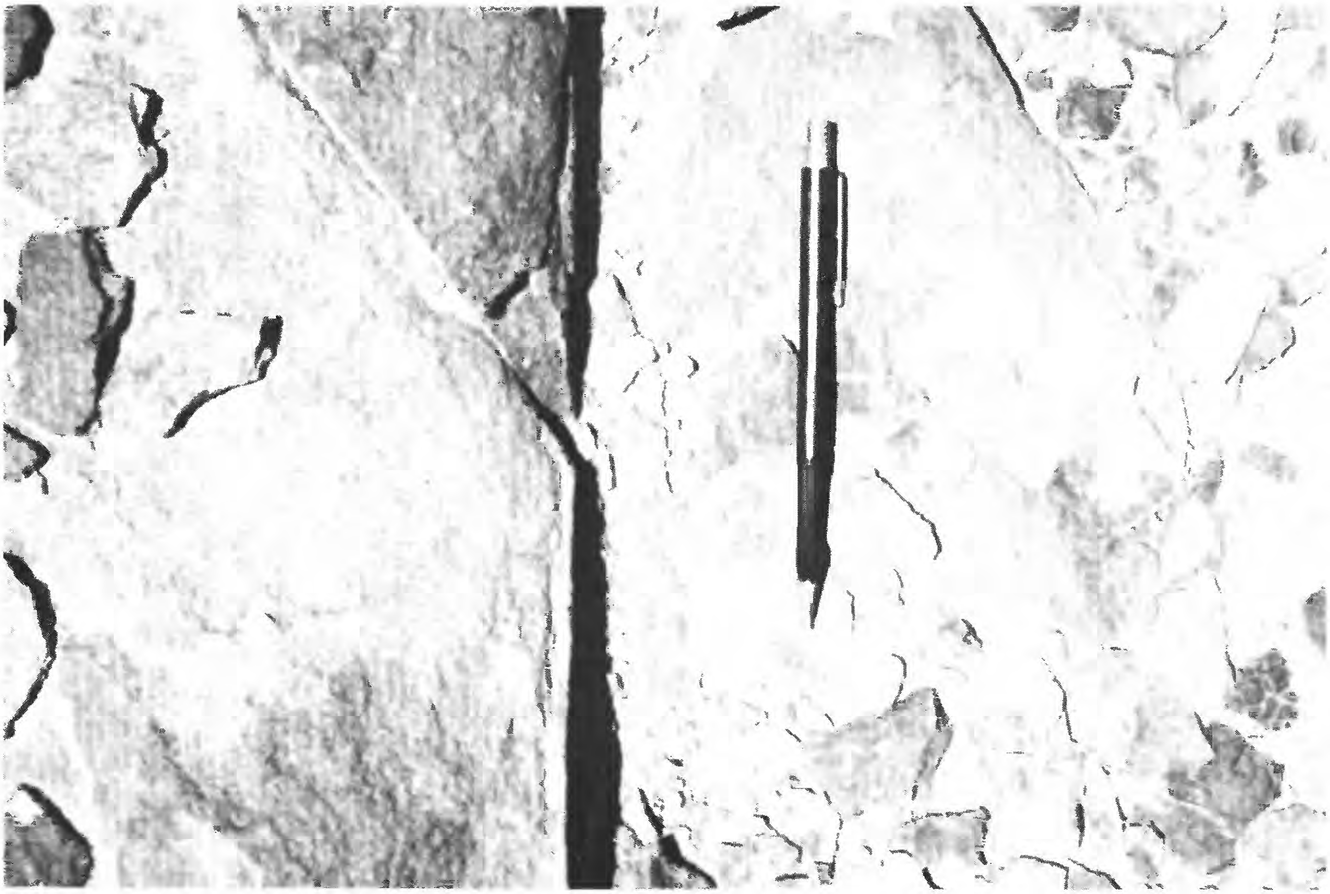

Figure 29. Horizontal bedding surface showing termination of F_2 extension joint (diagonal fracture, upper left) against dike-parallel joint (parallel to pencil) near dike of Pariette system (out of view to right), NW1/4 sec. 4, T. 9 S., R. 17 E. Plan view; top of photograph to northwest. Angular discordance between dike-parallel and F_2 joints is greatest in this area; strikes of two fractures shown are N. 43° W. and N. 80° W., respectively.

generally, maximum depths of emplacement—that is, depths to the bases of gilsonite dikes at their time of formation—can be estimated from structure-contour maps of present depths to the Mahogany zone (Smith, 1981) in conjunction with estimates of amounts of overburden removed from the region by erosion. Several factors complicate the exercise, chief among them being that almost all trace of the surface of maximum aggradation has been removed from the basin interior (Anders and others, 1992). The youngest beds preserved in the eastern Uinta basin are those of the Duchesne River Formation (plate 1*A*), locally of Oligocene age; hence most direct evidence of the nature and thickness of younger beds that may once have occupied the area has been lost. Estimates of the thickness of section removed from the interior of the Uinta basin thus are based mostly on such indirect methods as extrapolation of vitrinite reflectance profiles to estimated surface values, reconstruction of burial and thermal histories using time-temperature Lopatin-style techniques, and geologic inference. Explanation of these methods and discussion of the results obtained are given in Anders and others (1992), Fouch and others (1992), and Johnson and Nuccio (1993) .

Estimates of former overburden thickness are available for four areas within the region of the gilsonite dikes. Proceeding roughly from west to east, the first of these is the Pariette Bench oil field, near the Green River at the southeastern end of the dikes of the Pariette system. Pitman and others (1982) used projected thicknesses of stratigraphic units to estimate for this field that no more than about 1,000 m of overburden has been removed. Shale compaction data from a well in the same area led Sweeney and others (1987) to estimate a former overburden of 2,302–3,189 m, but this value seems high in light of their other reported overburden estimates for the region. Across the river to the southeast, Johnson and Nuccio (1993) plotted a vitrinite reflectance profile using samples from two wells (in secs. 7 and 8, T. 10 S., R. 20 E.) to estimate the position above the present-day land surface of the former pre-uplift erosion surface. Gaps in the depth distribution of the data made it difficult to recognize possible kinks in the profile, with consequent uncertainty in the slope of the upper part of the profile and thus the position of the surface intercept. Despite the inherent difficulties, the data nevertheless suggest that only modest amounts of overburden—700 m or less—have been removed. To the northeast is the Natural Buttes gas field, where reconstruction of the burial and thermal history of the rocks using a time-temperature Lopatin model led Pitman and others (1987) to conclude that about 1,000 m of Uinta and Duchesne River strata has been removed from the area. The Natural Buttes gas field occupies roughly the same area as the gilsonite dikes of the Ouray system. Finally, Johnson and Nuccio (1993) used a vitrinite reflectance profile from the Mid-America No. 1 Unit well near Bonanza (sec. 24, T. 9 S., R. 24 E.) to estimate that between 1,220 and 2,740 m of overburden has been removed; the well site is within the central part of the Cowboy-Bonanza system of dikes. Comparable values of from 1,410 to 1,802 m, each with an uncertainty of about 300 m, were obtained by Sweeney and others (1987) from shale compaction curves for three wells nearby to the north and northwest.

A different approach employed recently by Anders and others (1992) is based on the assumptions that (1) remnants of a late Eocene to early Oligocene erosion surface in the Piceance basin to the east and along the south flank of the Uinta Mountains to the north approximate the maximum regional surface of aggradation and (2) present elevations of those remnants can be used to estimate the elevation of the analogous surface, now removed, that probably once extended across the interior of the Uinta basin. Although post-Laramide collapse has tilted the erosion surface flanking the Uinta Mountains, its maximum elevation of about 3,050 m is similar to the present elevation of the erosion surface in the Piceance basin; thus 3,050 m was chosen as a regional average. Inferred amounts of overburden for two sites, one bordering the area of gilsonite dikes on the north and the other on the southwest, are 1,464 and 1,525 m, respectively. However, time-temperature modeling of the burial history using these values predicts higher hydrocarbon maturation values than those actually observed in samples from the two well sites. Adjustment of the data to bring the modeled maturation levels in conformity with those measured reduces the inferred amounts of eroded overburden to between 488 and 567 m (Anders and others, 1992).

Fouch and others (1992), in discussing these and similar data for other parts of the basin, noted that published estimates of overburden removed vary greatly, from about 300 to almost 3,350 m, but that values within the lower part of this range are more consistent with structural and stratigraphic reconstructions and with maturation patterns. Accordingly, we adopt 500–1,000 m as a conservative estimate of the amount of overburden removed from the region occupied by the gilsonite dikes since their time of emplacement. Together with information on current depths to the source-rock zone, these values suggest emplacement depths of 700–1,300 m for dikes of the Rainbow and Cowboy-Bonanza systems near the basin margin, increasing northwestward to as much as 2,500 m for dikes of the Fort Duchesne system in the basin interior. These estimates have been revised slightly downward from those we reported previously (Verbeek and Grout, 1992) in light of newly published information.

SOURCE-ROCK MATURITY AND THE MIGRATION OF GILSONITE

Oil in several fields coincident with or marginal to the outcrop area of gilsonite dikes is produced mainly from middle Eocene reservoir rocks too thermally immature to have generated it. Some authors, as discussed following, now view this as evidence for migration of oil from distant, deeper, more thermally mature source beds to the northwest. We here speculate that gilsonite, too, may have migrated appreciable distances laterally through the dike fracture system and within parts of the dike swarm may not have been derived from the Green River beds directly beneath. Though the distances involved need not be as great for the immature gilsonite as for the moderately mature oils, the available evidence nevertheless suggests that much of the gilsonite, particularly that in the southeastern part of the dike province, was derived from a downdip source.

The Red Wash field, bordering the area of the gilsonite dikes on the north, is a prime example of a field producing from migrated oil. Biomarker ratios for oils from this field indicate a thermal maturity equivalent to vitrinite reflectance (R_m) values of 0.7–0.8 percent (moderately mature), yet the Green River reservoir rocks are thermally immature at only 0.4–0.55 percent R_m (Anders and others, 1992). Osmond (1992) suggested that oil in the Red Wash field was derived from Green River beds in the deepest part of the basin and migrated southeastward more than 60 km to its present position. Chidsey and Laine (1992) suggested the same for the Pariette Bench field, citing as evidence the compositional similarity between oil produced from that field and from deeper, stratigraphically equivalent zones in the Altamont-Bluebell area to the north and northwest. Indeed, the Green River beds throughout the area occupied by dikes of the Ouray, Willow Creek, Rainbow, and Cowboy-Bonanza systems are thermally immature with respect to the oil-generation window as defined by Lopatin models (Osmond, 1992; Anders and others, 1992); thus the question arises as to the maturity of these same beds with respect to gilsonite.

Generation of gilsonite at very low levels of thermal maturity was suggested by Palacas and others (1989) and Monson and Parnell (1992) as one possible explanation for the presence of gilsonite dikes in shallow parts of the basin. Much support for this hypothesis comes from the gilsonite itself, which is highly aromatic (Cornelius, 1984; Khavari-Khorasani, 1984) and by any measure has experienced only a mild thermal history (Palacas and others, 1989; Hatcher and others, 1992). Biomarker ratios for gilsonite suggest a degree of thermal maturity equivalent to a vitrinite reflectance of only 0.5 percent (Anders and others, 1992, p. 74), compatible with derivation from source rocks too immature to have produced commercial quantities of conventional oil. Much of the upper Green River Formation in the area occupied by the dikes, however, has not reached even this level of maturity. The 0.5-percent R_m surface in the Island field, for example, is well below the Mahogany zone (Nuccio and others, 1992, fig. 2; Johnson and Nuccio, 1993), and thus well below the known or suspected bases of most gilsonite dikes, yet multiple dikes of the Ouray system crop out in the same area. A similar situation exists at the Pariette Bench field (Nuccio and others, 1992) and farther southeast near Bonanza (Johnson and Nuccio, 1993). Some degree of migration thus may be implied by the widespread occurrence of gilsonite dikes above source rocks that are either immature or at best only marginally mature with respect to gilsonite.

Migration of fluids within gilsonite dikes over map distances of 10 km or more seems plausible, if not likely, in view of dike structure and evolution as presented in this paper. Surface structures on the walls of the Cowboy dike near the White River (fig. 11), as noted previously, indicate southeastward propagation of the fracture zone, showing that the aqueous fluids originally expelled through these fractures were derived from a source to the northwest. Similar relations were documented along the same dike in nearby trenches. Though no one has yet studied the orientation of surface structures on a sufficient number of dikes to document the regional pattern of fluid transport within them, the great lengths of many dikes relative to their vertical extents suggest that the dominant flow paths were lateral. Some appreciation for possible migration distances can be gained from the dimensions of some of the larger dikes: 23 km in outcrop for the Cowboy dike (Pruitt, 1961), with a lengthy subsurface continuation to the northwest (E.V. Deshayes, quoted in Barb, 1944, p. 16); 13–18 km for several dikes of the Ouray and Willow Creek systems (Pruitt, 1961); and almost 30 km for the Pride-of-the-West dike and its southeastern continuation, the Rainbow dike. The Ouray–Willow Creek and Rainbow dike systems, if connected at depth as postulated by Crawford and Pruitt (1963), would form a single dike complex at least 65 km long along strike. Significant lateral transport of gilsonite is also compatible with evidence of high fluid pressures during expulsion of the gilsonite and with dike-fracture propagation by a hydraulic fracture mechanism, as discussed in the following section.

Migration of gilsonite from a downdip source and its emplacement in beds nearer the basin margin could explain two hitherto puzzling features of dike geology. First is that bitumen-rich beds of the Green River Formation, where observed in direct contact with gilsonite dikes in the Rainbow-Dragon area near the Colorado State line, show no evidence of depletion of their organic content near the dikes (Cashion, 1967). This is to be expected if the source beds are not those in direct contact with the dikes but instead are laterally equivalent beds in deeper, unexposed areas to the northwest. Second is the presence of several dikes stratigraphically below their source horizons, as exemplified by the Black Dragon dike. These too are to be expected if the gilsonite within them migrated southeastward, toward beds dipping gently in the opposite direction.

INTERPRETATION

The apparent uniformity in late Cenozoic fracture history between the eastern Uinta and Piceance basins suggests that one must look beyond the confines of a single basin for the cause of the deformation that gave rise to gilsonite dikes as one of its products. In this section we suggest that the gilsonite dikes are early products of the same period of regional extension that shortly thereafter resulted in formation of the F_2 joint set over a far wider area and somewhat later resulted in normal faults of small to moderate throw. Structures analogous to the gilsonite dikes, but not previously recognized as such, are present in the Piceance basin as well; these are discussed in a later section.

Gilsonite Dikes as Hydraulic Fractures

Notable features of the gilsonite dikes are their large size, their early age relative to other structures in the host rocks, the passage of aqueous fluids through the dike fracture network before its invasion by gilsonite, and the abundant evidence for forceful (fig. 13) rather than passive intrusion. The widespread distribution of sills (figs. 17, 19, 20), in particular, shows that fluid pressures during gilsonite injection frequently exceeded lithostatic load. Bleaching of the wallrock, coatings of limonite and calcite on dike walls, and deposition of chlorite in the adjacent sandstones all predate the gilsonite and record early fluid circulation; the presence of chlorite suggests derivation of the water from a magnesium-rich, probably dolomitic source rock (Monson and Parnell, 1992). From these relations, as well as the almost overwhelming evidence that the gilsonite itself was derived from oil shale, we conclude that the gilsonite dikes began as large-scale hydraulic extension fractures from overpressured hydrocarbon source beds in the Green River Formation (Verbeek and

Grout, 1992). Breaching of the overpressured zone by extensile failure and propagation of fracture zones through overlying beds for distances commonly exceeding 10 km laterally and 2 km vertically allowed the escape of formation waters on a possibly large scale.

Stress State during Dike Formation

Several inferences about the stress state during formation of the gilsonite dikes are germane to an understanding of their origin. In the discussion that follows, σ_1 and σ_3 refer to the maximum and minimum principal stresses, respectively; σ_{hmax} and σ_{hmin} to the maximum and minimum stresses *in the horizontal plane*, respectively; and σ_v to the overburden stress acting in the vertical direction. Note for a vertical extension fracture that $\sigma_{hmin}=\sigma_3$ but that σ_{hmax} does not necessarily equal σ_1; the latter need not be horizontal. In order of increasing stress magnitude, $\sigma_3=\sigma_{hmin}<\sigma_2\leq\sigma_{hmax}\leq\sigma_1$.

1. Given that the rocks were unfractured before formation of the hydraulic dike fractures, the dikes may be taken as valid indicators of regional stress directions during the hydraulic fracturing process. The dikes are vertical and show strong northwest-preferred orientation, and surface structures on their walls confirm that they originated as extension fractures. Regional (remote) σ_3 during breaching of the overpressured zone thus was horizontal and directed northeast, perpendicular to the planes of the dikes. With respect to horizontal stresses, trajectories of regional σ_{hmax} and σ_{hmin} ($=\sigma_3$) during hydraulic fracture formation were west-northwest to northwest and north-northeast to northeast, respectively parallel and perpendicular to the outcrop traces (plate 1*A*) of the dikes. Similar but not identical stress directions were later necessary for the development of the F_2 set of joints (plate 1*B*).

2. Spatial restriction of dike-parallel joints to local zones of stress amplification adjacent to dikes, and elsewhere to rare beds of exceptional brittleness, implies that regional stress differences (remote $\sigma_1-\sigma_3$) during dike-fracture formation remained sufficiently low that most strata in the interdike areas could not fracture. Nevertheless, the ratio of maximum to minimum horizontal stress (regional $\sigma_{hmax}/\sigma_{hmin}$) at this time must have been sufficiently greater than unity to allow the strong preferred dike trend to develop.

3. The common presence of gilsonite sills in association with the dikes shows that the original hydraulic fractures were forcibly opened to make room for the gilsonite that now fills them. Fluid pressures within the dikes exceeded both σ_v and σ_3 in the host rocks during invasion of the gilsonite. That the walls of both sills and dikes commonly are coated with limonite and bordered by altered rock extends the same conclusion to the aqueous fluids that formerly migrated through the dike fractures. High fluid pressures apparently were maintained or restored in the source beds during at least the initial stages of hydrocarbon generation and migration.

Suggested Evolution

The gilsonite dikes of the eastern Uinta basin are one of several types of structures that collectively record the regional late Cenozoic stress history of the northern and northeastern margins of the Colorado Plateau in western Colorado and eastern Utah. Events leading to the formation of the gilsonite dikes and related structures are summarized below. Though in this discussion we make repeated reference to fluid overpressures within the Green River Formation, interpretation of specific mechanisms by which such pressures could have formed and been maintained is beyond the scope of this paper. We note only in passing that the Green River source beds are rich in organic matter, that low to exceedingly low matrix permeability is characteristic not only of these beds but also of much of the overlying section, and that the overlying clastic sediments were deposited rapidly. Poorly drained conditions favoring both the buildup and maintenance of high pore pressure likely prevailed for much of the early history of these strata. The interplay between elevated fluid pressures in the source rocks and tectonic stress imposed on these same rocks is key to the interpretation of the gilsonite dikes.

The earliest postdepositional macroscopic structures known to us within the Green River Formation are clastic dikes and sills derived from laterally extensive beds of ash-fall tuff interstratified with the lacustrine oil shale. These early intrusions are common within the Piceance basin (Grout and Verbeek, 1982) and have been found also within parts of the eastern Uinta basin (Cashion, 1967). Deformation of the wallrock adjacent to many of the tuffaceous dikes (Grout and Verbeek, 1982) shows that the oil shale was stiff and capable of fracture but not yet fully consolidated at the time of dike intrusion. Some dikes terminate upward in small laccolithlike bodies from which sills extend laterally. An early history of fluid overpressure within the Green River Formation is implied. No hydrocarbons have been reported from these dikes, and it is likely that temperatures and burial depths were still too low for their generation. Lack of any evidence of preferred strike of the dikes (Cashion, 1967, p. 19; Grout and Verbeek, 1982) indicates that horizontal stresses at this time were almost isotropic.

The next event in the structural record was formation of the F_1 joint set. These joints, though documented at hundreds of localities within the Uinta and Piceance basins, show marked variation in geographic prominence and are all but absent from the area of the gilsonite dikes. Though we know from indirect evidence that the F_1 event almost certainly predated the gilsonite dikes, to date we have nowhere found F_1 joints in direct association with them. In any case, stress orientations during F_1 time were incompatible with gilsonite-dike formation, and a relation between the two is unlikely.[1]

Following the F_1 event there occurred a prolonged period of regional extension whose structural products, in order of formation, include gilsonite dikes, F_2 joints, and small normal faults. Stress directions during this time varied within relatively narrow limits, as reflected in the west-northwest to northwest trend common to all these structures. Formation of the large hydraulic fractures later to become the gilsonite dikes was aided both by decrease in the magnitude of regional σ_{hmin} during the initial increments of tectonic extension and by elevated fluid pressures in the source rocks. Both conditions would have much enhanced the likelihood of extensile failure of the rock. That limonite and calcite were the first materials deposited on the fractures so produced suggests that the liquid precursor to gilsonite had not yet formed in abundance; instead we infer that overpressured formation waters were the first fluids to be expelled through the fracture network. The large lateral and vertical extents of the linear fracture zones attest to the large fluid volumes and pressures, and possibly large migration distances involved. The lack of any literature reference to wallrock clasts from beds far above implies that sinking of clasts through wide, water-filled fissures did not occur and thus that fracture openings during this period likely remained narrow.

Breaching of the overpressured source beds and partial loss of fluids through an extensive fracture system likely led to significant fluid-pressure declines within parts of the Green River Formation. Nevertheless, the evident force with which liquid bitumen subsequently was expelled into the fractures to form, on solidification, the gilsonite dikes and sills shows that high fluid pressures were maintained or restored during at least the early stages of hydrocarbon generation within the source beds. Most of the host rocks by this time were well lithified, with the exception of some mudstone beds of the upper part of the Uinta Formation that deformed plastically during injection of the viscous gilsonite. Dilation of the fractures to their present widths probably occurred during this phase rather than before.

At this point in the structural evolution, most of the Green River and younger strata still remained unbroken. The F_1 set had already formed but was only weakly expressed in the dike area, and the dikes themselves, though lengthy, occupied narrow zones between which lay wide expanses of unfractured rock. During continued regional extension, however, almost all of the strata were stressed to the point of failure, and the F_2 set of joints formed across a large region encompassing almost the whole of the Uinta basin and the northern and central Piceance basin. Counterclockwise rotation of the stress field since the time of gilsonite intrusion (Verbeek and Grout, 1986) produced the observed geometry of the joints to the dikes: near parallelism in some places and slight to moderate angular discordance in others, the joints having generally more westerly strikes (fig. 22). The lack of obvious F_2 joints as a component of pencillated structure within dikes and sills (fig. 27) suggests that the gilsonite during F_2 time had not yet hardened to a brittle solid and emphasizes that both the dikes and the joints are related products of the same deformation, probably not far removed from each other in time. Nevertheless we stress the genetic difference between them: the F_2 joints, unlike the dikes, are of regional (extrabasinal) distribution and are present over large areas, not only in the Tertiary basin rocks but also in pre-basin Cretaceous rocks stratigraphically far beneath the gilsonite source beds (Verbeek and Grout, 1984; Grout and Verbeek, 1985, 1992). The different conditions of formation are reflected in the different properties of the dike walls and joint surfaces. There is nothing in the geometry, areal and stratigraphic distribution, or postulated evolution of the F_2 joints to suggest that they, like the gilsonite dikes, are hydraulic fractures.

By analogy to the Piceance basin, the west-northwest- to northwest-striking normal faults shown by Cashion (1967) in the southeastern Uinta basin probably are late products of the same regional extension discussed above. Most faults of this trend in the Green River and younger rocks of the central Piceance basin represent zones of F_2 joints reactivated in shear above presumed normal faults in deeper strata. Cashion's description of the faults in the southeastern Uinta basin as vertical, or almost so, suggests that they formed in a similar manner.

ANALOGOUS STRUCTURES

Vertical, sheetlike bodies of calcite, similar in dimension to the gilsonite dikes, are present in the central part of the northern Piceance basin. Generally they have been mapped as faults, but we argue here that the faulting was secondary and that these bodies, like the gilsonite dikes farther west, were created as large hydraulic fractures due to fluid overpressures in the Green River Formation. The calcite dikes strike N. 60°–70° W., very nearly parallel to the F_2 joints and small normal faults that cut the same rocks. Most are 0.1–0.8 m thick and are filled with calcite exhibiting well-developed comb layering that resulted

[1]The converse may apply, however, to other hydrocarbon dikes farther west in the basin. For example, results of recent field studies (E.R. Verbeek and M.A. Grout, unpub. data, 1992) indicate that the tabbyite dike in Tabby Canyon south of Duchesne may bear the same relation to the F_1 period of fracture as the gilsonite dikes to the F_2 event. The ozokerite dikes near Soldier Summit may be of similar origin but have not yet been re-examined.

from crystal growth in successive layers from the dike walls inward. Continued precipitation ultimately filled most of the void space, but lenticular vugs lined with terminated calcite crystals are present locally in the central parts of some dikes. Stratigraphic offset across some of these features is nil, but others were reactivated as high-angle normal faults in post-F_2 time and exhibit throws of 10–60 m. The fault movements brecciated some of the previously deposited comb-layered calcite, angular clasts of which commonly are embedded in massive calcite of later deposition. The calcite fill in some of these faults is texturally complex and indicative of multiple episodes of deposition, brecciation, and recementation. Also present locally in the Green River Formation are calcite sills as much as several centimeters thick; these provide additional evidence of overpressured conditions but have not been much studied.

Hydrocarbons associated with the calcite dikes are inconspicuous but present in minor amount. Residues of oil on calcite crystals lining lenticular vugs within the dikes are common, and small veins of solid bitumen associated with calcite and brecciated wallrock have been reported from a few areas (O'Sullivan and Ging, 1987). Under ultraviolet radiation much of the calcite luminesces in various pastel colors (Verbeek, 1982) that almost certainly are due to organic activators or inclusions. The paucity of vein hydrocarbons in the Piceance basin in contrast to their evident abundance in the Uinta basin reflects differences in hydrocarbon maturation between the two regions, attributed by Franczyk and others (1989) to the shallow burial depth of even the lowest oil-shale beds of the Green River Formation in the Piceance basin. Apart from conspicuous differences in filling material, however, the calcite dikes and the gilsonite dikes seem to be mechanically analogous structures. Both are of comparable dimension, orientation, and age relative to regional joint sets and probably were produced by similar processes.

CONCLUSIONS

The gilsonite dikes of the eastern Uinta basin originated as large hydraulic extension fractures from overpressured, hydrocarbon-rich source beds in the Green River Formation during the early stages of post-Laramide regional tectonic extension. Bleaching of wallrock and deposition of limonite and calcite as early deposits on dike walls suggest that significant quantities of formation water were expelled through the fracture system, much of which was subsequently intruded by the viscous bitumen that solidified to gilsonite. Emplacement depths are estimated at 700–2,500 m. The widespread presence of gilsonite sills injected along bedding shows that fluid pressures at the time of injection frequently exceeded lithostatic load. Continuing tectonic extension later gave rise to a regional set of joints and ultimately to normal faults of small to moderate throw; both types of structures trend at low to moderate angles to the dikes but are present over a far wider area and a much greater thickness of section. The deformation thus progressed with time from local hydraulic extension fracture (gilsonite dikes) through regional nonhydraulic extension fracture (joints) to minor shear failure (normal faults at depth, reactivated joints nearer the surface) of the basin strata.

The structure of individual gilsonite dikes is strongly dependent on rock type. Dikes in sandstone generally are large, continuous, tabular bodies flanked by narrow zones of short, dike-parallel fractures; their geometry is similar to that of the igneous dikes and associated fracture zones documented by Delaney and others (1986) in similar rocks elsewhere. Numerous gilsonite dikes in sandstone are sufficiently thick, 0.5–5.5 m, to have been mined. Dikes intruded into weakly indurated mudstone, in contrast, form complex anastomosing networks of discontinuous dikelets too thin for exploitation.

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