Depositional History of Triassic Rocks in the Area of the Powder River Basin, Northeastern Wyoming and Southeastern Montana

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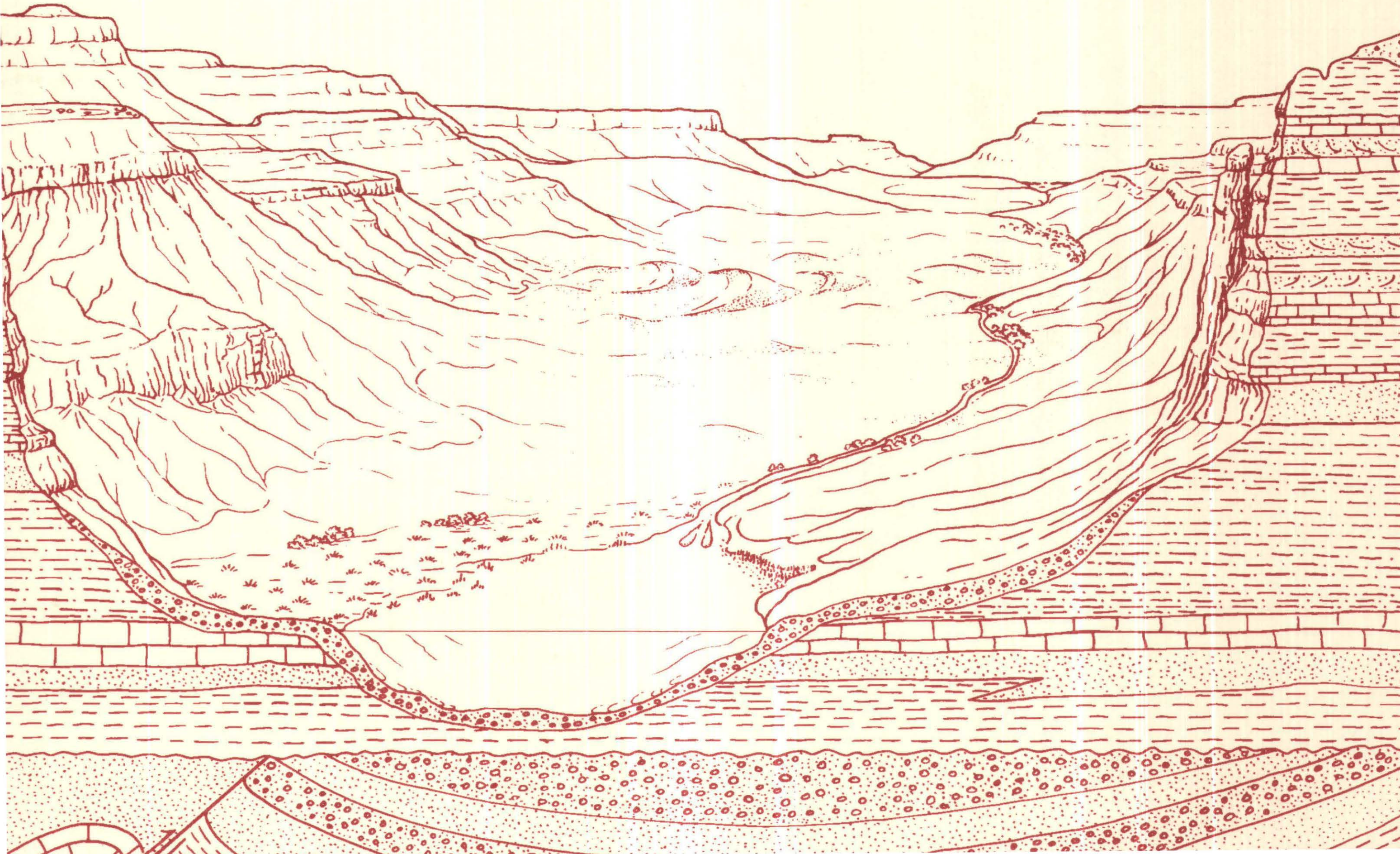

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Chapter P

Depositional History of Triassic Rocks in the Area of the Powder River Basin, Northeastern Wyoming and Southeastern Montana

By EDWARD A. JOHNSON

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 1917

EVOLUTION OF SEDIMENTARY BASINS—POWDER RIVER BASIN

U.S. DEPARTMENT OF THE INTERIOR
BRUCE BABBIT, Secretary

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Depositional History of Triassic Rocks in the Area of the Powder River Basin, Northeastern Wyoming and Southeastern Montana

By Edward A. Johnson

Abstract

Throughout the Early Triassic Epoch, the area that is now the Powder River Basin existed east of a large marine basin centered in southeastern Idaho. At the beginning of the Triassic silty sediments sourced from the east were being deposited under low-energy conditions in a nonmarine environment. During the earliest Early Triassic this setting was interrupted by a rapid marine transgression from the west, and carbonate and evaporite minerals were deposited in shallow water over a broad area that included much of the present Powder River Basin. Following the westward withdrawal of this sea, a flood of silty sediments prograded westward and were deposited on a flat surface under mixed marginal- and shallow-marine conditions. These deposits now form the bulk of the Lower Triassic red beds in the region. At the end of this depositional phase, a change in the source area, coupled with an increase in depositional energy, resulted in sandy sediments accumulating in a paralic environment. Another rapid marine transgression occurred during the late Early Triassic, and limestone was deposited in a shallow sea over a broad area; however, only the eastern part of the Powder River Basin was affected by this invasion. A relatively short period of nondeposition followed the westward withdrawal of this sea. In the latest Early Triassic, sandy sediments prograded westward and were deposited in a shallow-marine environment.

During the middle part of the Triassic a major erosional event occurred throughout the Western Interior of North America, and most, if not all, of Middle Triassic history is evidently unrepresented in the rock record. The large marine basin that had dominated the Western Interior during the Early Triassic did not survive this erosional event and was replaced during Late Triassic time by a broad region characterized by nonmarine deposition. In the area of the Powder River Basin, the erosion surface was covered by westward-prograding mud, silt, and sand deposited under fluvial conditions. Nonmarine deposition probably continued in the northern Western Interior until the end of the Triassic, but in the area of the Powder River Basin Jurassic erosion has removed this part of the rock record.

INTRODUCTION

As outlined by the contact of Cretaceous and Tertiary strata, the Powder River Basin covers about 22,000 mi^2 in northeastern Wyoming and southeastern Montana (fig. 1). Basement-cored uplifts border the basin on three sides: on the west are the Bighorn Mountains, to the south is the northern end of the Laramie Range and the Hartville Uplift, and to the east are the Black Hills. The northern limit of the basin is defined by the Miles City Arch, a buried, northwest-trending positive element on the Precambrian basement. This structure separates the Powder River Basin from the Williston Basin in eastern Montana. Triassic rocks are exposed in each of the three surrounding uplifts. The Powder River Basin was formed mostly during the early Tertiary and is on the eastern edge of the Laramide structural province.

BRIEF HISTORY OF THE TRIASSIC PERIOD IN THE WESTERN UNITED STATES

General Statement

Throughout the Triassic, all of the present-day continents were positioned together in the Pangean supercontinent (Smith and Briden, 1977). Just prior to the Late Permian or the Early Triassic, subduction is postulated to have occurred somewhere off the west coast of what is now North America along a west-dipping Benioff zone as

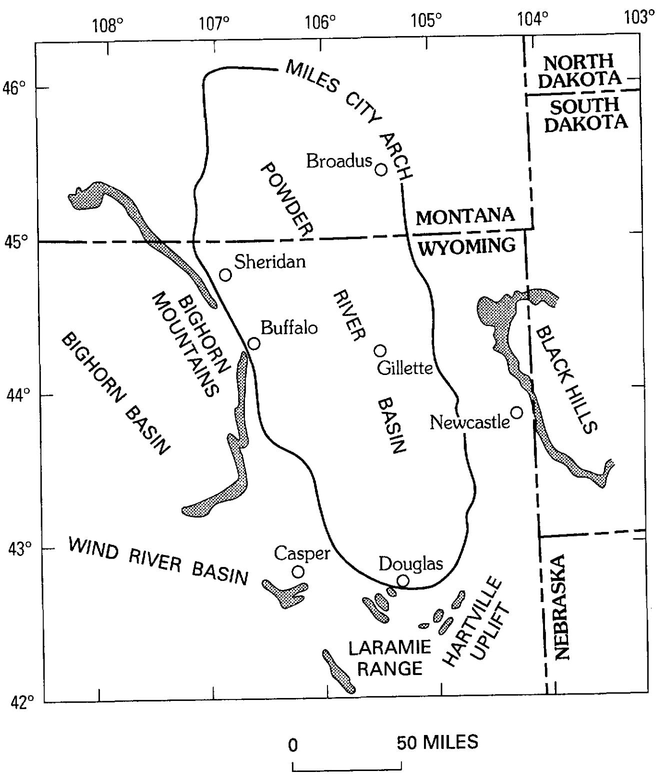


Figure 1. Location of Powder River Basin (shown by outline). Shading indicates areas of exposed Triassic rocks.

an east-moving island arc overrode oceanic lithosphere along the western edge of the Pangean plate (Speed, 1979). The island arc eventually collided with the passive western margin of the Pangean craton during the Late Permian or the Early Triassic, causing the Sonoma orogeny; during this collision, significant amounts of terrane were welded to the craton. As a result of this collision, the Benioff zone apparently reversed polarity and became east dipping (Condie, 1989), and subduction resumed as east-moving oceanic lithosphere moved under the western edge of the craton. This change in subduction direction resulted in the formation of an Andean-type continental margin along this part of the western edge of Pangea during the Late Triassic. In contrast to this model, Burchfiel and Davis (1981) proposed that closure of the initial basin separating the island arc from the craton was accomplished by eastward subduction; thus, their model does not require a reversal of subduction direction following collision. They also postulated that during the earliest Triassic a significant left-lateral transform fault formed a major plate boundary in what is now the southwestern United States.

Triassic sedimentary rocks are exposed as far south as the central parts of Arizona and New Mexico, and western Texas; as far east as central North Dakota, western South Dakota, northwestern Nebraska, and the eastern part of Colorado; and into Canada to the north (McKee and others, 1959; MacLachlan, 1972). On the west, Triassic sedimentary rocks have been identified in roof pendants in the Sierra Nevada of southern California (Saleeby and others, 1978). In general, thick basinal marine deposits thin eastward through shelf units into nonmarine deposits. Triassic rocks in the Western Interior are widely distributed and usually well exposed. As a result, studies of Triassic rocks in this part of the western United States abound. Interpreting Triassic history west of the eastern Great Basin is more difficult because postdepositional tectonism has disrupted the original distribution of Triassic deposits to such an extent that reconstructing paleogeography is rather speculative.

Early Triassic marine fossils have been identified in the Western Interior from the Dinwoody and Thaynes Formations and from the western part of the Moenkopi Formation (MacLachlan, 1972). In the Rocky Mountains, a broad, westward-sloping, stable cratonic platform existed during the Early Triassic on which red beds were deposited as an eastward-tapering wedge (fig. 2). In general, the grain size of the red beds increases somewhat from north to south. The Red Peak and Triassic part of the Spearfish Formations were deposited on the northern part of this platform in a mixed marine and nonmarine environment. On the southern part of the platform, the Moenkopi Formation was deposited in a coastal-plain setting that included fluvial, tidal flat, and shallow-marine environments (Blakey, 1989). The source areas for clastic detritus in the red beds were the exposed craton far to the east, the Ancestral Rocky Mountains and the Uncompahgre Uplift, and smaller uplifts in Arizona, New Mexico, and Montana. The eastern side of the Cordilleran miogeocline was west of the Rocky Mountains in what is now the eastern Great Basin. The line separating the marine shelf from the platform probably trended northeast through what is now central Utah. Lower Triassic rocks deposited in this sea are represented by marine carbonate and shale in the Dinwoody and Thaynes Formations. McKee and others (1959) reported as much as 5,000 ft of Lower Triassic marine rocks in southeastern Idaho that represent the most rapidly subsiding part of the basin. Through a series of rapid transgressions and regressions, extensions of this sea to the east deposited tongues of carbonates and evaporites, now interbedded with the red beds, as far east as the Black Hills of South Dakota. The regional interfingering of marine and nonmarine rocks is particularly well displayed in the Moenkopi where carbonate increases westward to the extent that near Las Vegas, Nevada, it represents more than 50 percent of the formation (Blakey, 1989). The final phase of marine sedimentation in this mostly Paleozoic age Cordilleran miogeocline occurred during the Early Triassic.

Middle Triassic deposits are, for the most part, missing from the Western Interior, and a major unconformity commonly separates Upper and Lower Triassic strata.

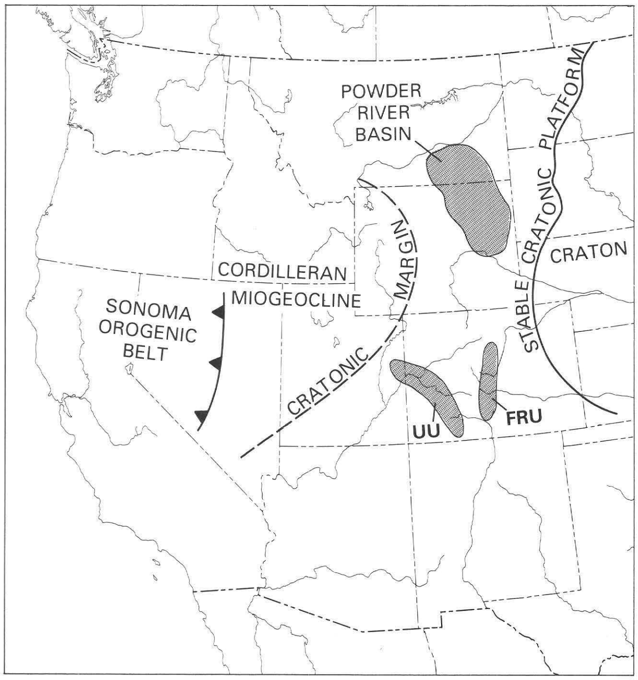


Figure 2. Generalized paleogeography of western United States during the Early Triassic. FRU, Front Range Uplift; UU, Uncompahgre Uplift. Modified from Blakey (1989), Peterson and Smith (1986), and Carr and Paull (1983).

Most workers (for example, Lucas and others, 1985; Blakey, 1989), however, consider the uppermost part of the Moenkopi to be earliest Middle Triassic in some areas.

Late Triassic fossils in the Western Interior are known from the Popo Agie Formation in the northern Rocky Mountains and from the Chinle Formation on the Colorado Plateau (MacLachlan, 1972). Upper Triassic deposits in the Western Interior are mostly continental in origin, and they extend farther to the southeast and less far to the northeast than do Lower Triassic deposits. The source areas for the detritus in these deposits were probably the same as those during the Early Triassic; however, unlike the Early Triassic, individual depocenters existed in the Western Interior during Late Triassic, as indicated by isopach maps of these deposits (MacLachlan, 1972). The Popo Agie and Jelm Formations of the northern Rocky Mountains and the Chinle Formation of the Colorado Plateau were deposited in fluvial, lacustrine, and eolian environments.

Conditions during the Triassic in the area west of the eastern Great Basin can be only partly reconstructed. Evidently, during much of Triassic time the marine shelf of the Cordilleran miogeocline extended west into central Nevada. The Triassic section in this area is as thick as 20,000 ft (McKee and others, 1959), and fossils dated from latest Early to Late Triassic have been reported. According to Speed and Silberling (1989), these strata record a complicated history of marine transgressions and regressions separated by periods of erosion. The resulting deposits contain a great variety of carbonate and clastic rocks, some as coarse as conglomerate. Somewhere west of this shelf a deeper part of the basin might have existed through at least part of the Triassic, but evidence for it is sparse. The early history of the western shelf was dramatically affected by the Permian to Early Triassic Sonoma orogeny that occurred when an island arc and inboard accretionary wedge collided with the craton. During this event it is postulated that in central Nevada the oceanic sediments of this accretionary wedge were thrust eastward as much as 60 mi over precollision rocks (the Golconda allochthon). Postthrusting Triassic units were thus deposited over this terrane. The history of Triassic deposition west of the eastern Great Basin differs from that to the east in two basic ways. First, certain Triassic deposits west of the eastern Great Basin contain significant amounts of volcanic and volcanogenic sedimentary rocks. This volcanism evidently resulted first from the initial collision and subsequently from the development of arc magmatism along this part of the western margin of the craton. Evidence of this magmatism is present in the southern Sierra Nevada where certain plutons have been dated as Late Triassic in age. Second, unlike the Triassic history to the east, the Middle Triassic is represented in the rock record. Evidently, when marine water withdrew from the eastern part of the miogeocline in the late Early Triassic, marine conditions continued sporadically in the western part of the basin through Middle and most of Late Triassic time. The shoreline during this time of what was left of the miogeocline must have been somewhere in Nevada. It was not until the latest Late Triassic that marine waters finally withdrew from the western part of the basin, signaling the final chapter of the Cordilleran miogeocline.

Paleolatitude and Climate

Maps in Parrish and Peterson (1988) show the area of the Powder River Basin to have been at about 15° N. during the Early Triassic and at about 23° N. during the Late Triassic. Parrish and Peterson reported that the Triassic climate in the western United States was probably humid with seasonal precipitation and occasional periods of drying; they compared this climate to that of present-day southwest India (see also Dubiel and others, 1991). Peterson (1988) noted the relative abundance of eolianite in Pennsylvanian, Permian, and Jurassic deposits as compared to deposits of the Triassic, and he attributed this to a wetter climate during the Triassic. In southeastern Wyoming, eolianites are present in the Upper Triassic Jelm Formation, and Peterson (1988) reported that field measurements indicate a northwesterly wind direction during this deposition. Based on global circulation models, Parrish and Peterson (1988) believed that Early Triassic winds were out of the northeast.

Early Triassic Depositional Setting on the Eastern Side of the Middle Western Interior

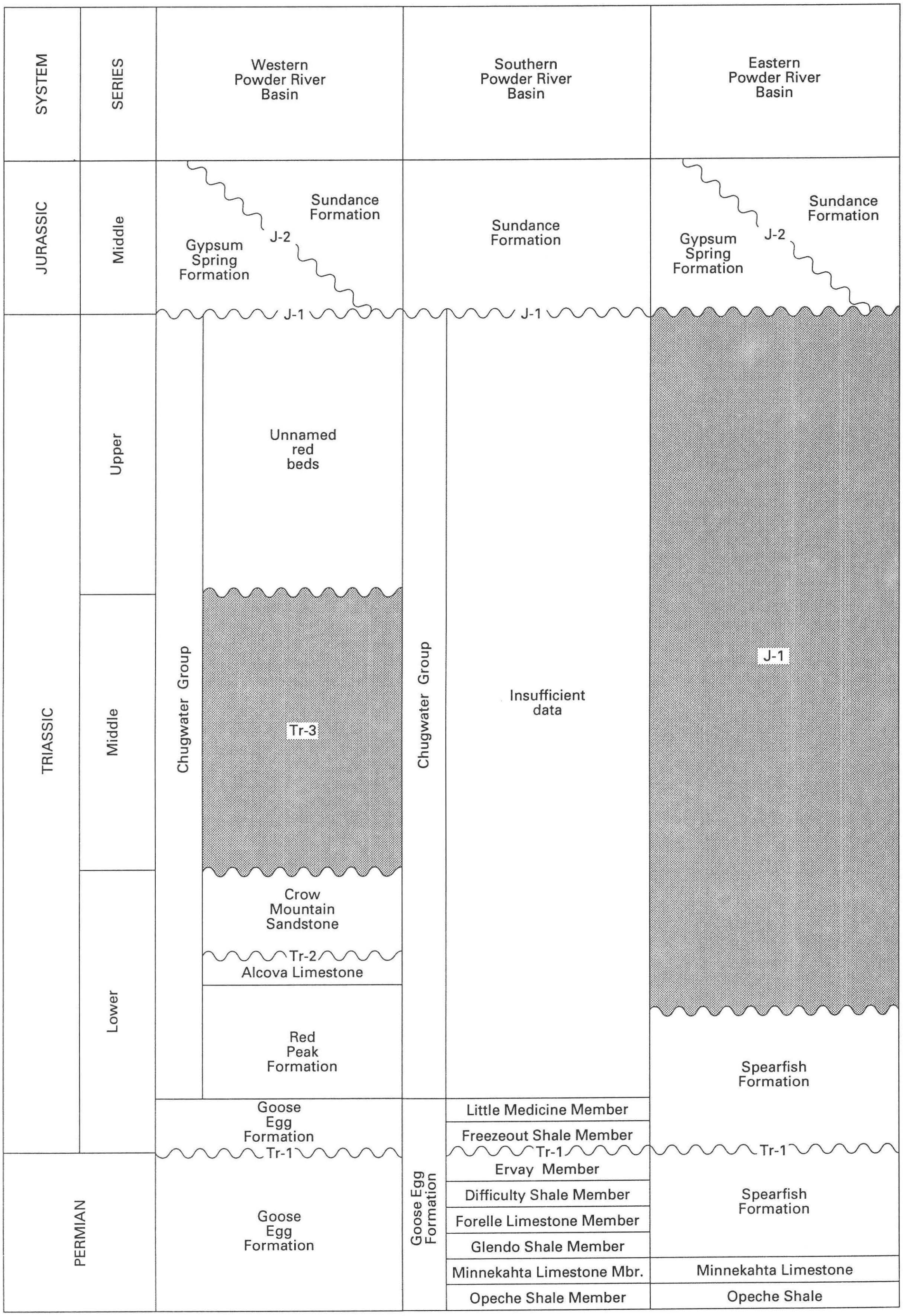
During the Late Permian through Early Triassic, the eastern side of the middle Western Interior including what is now central and eastern Wyoming, and southeastern Montana was a region of low relief, at, or just above, sea level, that sloped gently toward the west. The eastern shoreline of the Cordilleran miogeocline was somewhere to the west, and the exposed craton, also of low relief, was to the east. To the south in central Colorado, the Ancestral Rocky Mountains were exposed, and smaller uplifts possibly were present to the north in central Montana. Other than slow regional subsidence, it was a time of relative tectonic stability. The exact depositional conditions on this vast plain are somewhat enigmatic; it is possibly that no modern analog exists. What is known is that for millions of years uniform, low-energy deposition of mainly silt sized detritus occurred over thousands of square miles resulting in a colorful sequence of red beds in this area of the Western Interior. Four times during the Late Permian through Early Triassic marine transgressions from the miogeocline extended east onto this vast plain. Carbonates and evaporites were deposited during these relatively short events and are present as distinct units interbedded with the red beds. In central Wyoming the sequence of rocks representing this period of red-bed deposition interrupted by marine invasions is known as the Permian and Triassic Goose Egg Formation. The transition from Permian to Triassic time thus occurred during Goose Egg deposition. Following this accumulation, continued deposition of red beds created the bulk of the present-day red beds in what is now the Lower Triassic Red Peak Formation of the Chugwater Group.

STRATIGRAPHIC FRAMEWORK

The stratigraphic nomenclature of Triassic rocks in the area of the Powder River Basin as used in this report is shown on figure 3. The first mention of Triassic rocks in eastern Wyoming was by Knight (1902), who referred to them as the "Laramie Plains red beds." Darton (1904) formally named the Chugwater Formation for exposures along Chugwater Creek near Iron Mountain in the Laramie Range. As originally defined, the Chugwater Formation included all of the strata between the Tensleep Sandstone (Pennsylvanian) and the Sundance Formation (Jurassic). Later, in the Bighorn Mountains, Darton (1906) assigned the lower part of his original Chugwater to the Embar Formation, an early name that included what are now known as the Park City (Permian) and Dinwoody (Triassic) Formations of central and western Wyoming. Near the northwestern end of the Laramie Range, rocks equivalent to the Embar were later named the Goose Egg Formation, but the name Embar persisted in the northern Bighorn Mountains, even though Thomas (1934) had abandoned the term in central Wyoming in favor of the names Phosphoria (Permian) and Dinwoody. This resistance to change was because geologists working in the Bighorn Mountains could not divide the rocks into Phosphoria and Dinwoody, or identify individual members of the Goose Egg with any certainty. To get around this problem, some geologists (for example, Hose, 1955; Maple, 1959) used the term "gypsum and red shale sequence" (or vice versa) to describe these rocks. The Chugwater Formation was subdivided in the southern Absaroka Range by Love (1939) into the Red Peak, Crow Mountain, Popo Agie, and Gypsum Spring Members, in ascending order. The Gypsum Spring later proved to be Middle Jurassic in age (Love, Tourtelot, and others, 1945). Branson and Branson (1941) raised the Chugwater to group status and raised Love's members to formations. In addition, they included in the Chugwater the Dinwoody, and added the Nugget Sandstone (then called the Wyopo Formation) and strata they termed the Alcova Dolomite. The Nugget was later demonstrated to be Jurassic in age (Pipiringos and O'Sullivan, 1978). Pipiringos (1968) reaffirmed group status for the Chugwater, retained the Red Peak, Alcova Limestone, Crow Mountain, and Popo Agie Formations, and added the Jelm Formation. In the Laramie Range, Darton (1904) reported the average thickness of the Chugwater to be 1,250 ft. From here, the group thins toward the Hartville Uplift (Denson and Botinelly, 1949) and, evidently, also thins toward the west and north. In the area of the Powder River Basin, the group has a maximum thickness of about 800 ft in the central Bighorn Mountains (Mapel, 1959). In this area reported thicknesses of the group vary because of erosional beveling of its top, but in general the unit thins toward the north and east. The Chugwater overlays the Goose Egg and is overlain by either the J–1 unconformity and the Gypsum Spring or the J–2 unconformity and the Sundance. The depositional environments of the Chugwater become more terrestrial both upsection and laterally toward the east.

Darton (1899) introduced the term Spearfish Formation for red beds he observed below the Jurassic Sundance Formation in the Black Hills. The definition of the Spearfish was further refined by Darton (1901) as those rocks between the Minnekahta Limestone (Permian) and the Sundance; Darton also implied that exposures of the unit near the town of Spearfish, South Dakota, could serve as the type locality. Darton (1904, 1906) realized that the

Figure 3 (facing page). Generalized stratigraphy of Triassic and adjacent rocks in the vicinity of the Powder River Basin. Tr–1, Tr–2, Tr–3, J–1, and J–2 represent unconformities. The Sundance Formation sits with J–2 unconformity on the unnamed red beds and Spearfish Formation in some areas. Modified from Pipiringos (1968), Boyd and Maughan (1973), and Pipiringos and O'Sullivan (1978).


SYSTEM
SERIES
Western Powder River Basin
Southern Powder River Basin
Eastern Powder River Basin
JURASSIC
Middle
Gypsum Spring Formation
J-2
Sundance Formation
Sundance Formation
Gypsum Spring Formation
J-2
Sundance Formation
J-1
J-1
TRIASSIC
Upper
Middle
Lower
Chugwater Group
Unnamed red beds
Tr-3
Crow Mountain Sandstone
Tr-2
Alcova Limestone
Red Peak Formation
Chugwater Group
Insufficient data
J-1
Spearfish Formation
Goose Egg Formation
Tr-1
Little Medicine Member
Freezeout Shale Member
Tr-1
Tr-1
PERMIAN
Goose Egg Formation
Goose Egg Formation
Ervay Member
Difficulty Shale Member
Forelle Limestone Member
Glendo Shale Member
Minnekahta Limestone Mbr.
Opeche Shale Member
Spearfish Formation
Minnekahta Limestone
Opeche Shale

Spearfish and Chugwater were more or less equivalent. As originally defined, the Spearfish contained a gypsum unit at the top that was later identified by Imlay (1947) as the Gypsum Spring Formation. Strata equivalent to the upper part of the Goose Egg are now recognized in the lower part of the Spearfish.

Several major unconformities have been recognized within the Triassic sequence in eastern Wyoming (Pipiringos and O'Sullivan, 1978). The Tr–1 unconformity separates Permian rocks from Triassic rocks throughout the Western Interior, and its presence in eastern Wyoming will be discussed later in this report. The Tr–2 unconformity is above the Alcova Limestone, and the Tr–3 unconformity is above the Crow Mountain.

DEPOSITIONAL PACKAGES AND EROSIONAL EVENTS

Late Permian and Early Triassic Transgressive-Regressive Cycles

Between the earliest Late Permian and the earliest Early Triassic, several eastward transgressions of marine water from the Cordilleran miogeocline of westernmost Wyoming inundated the generally nonmarine setting of central and eastern Wyoming. As a result, tongues of marine carbonate and evaporite rocks from the west now interfinger with terrestrial clastic progradational wedges from the east, creating the Permian and Triassic Goose Egg Formation.

Interbedded Marine-Shelf Carbonate and Evaporite Rocks and Nonmarine (?) Red Beds

Burk and Thomas (1956) named the Goose Egg Formation for exposures of rocks near the Goose Egg Post Office 11 mi southwest of Casper, Wyoming. As now defined, the Goose Egg consists of the following eight members, in ascending order: Opeche Shale, Minnekahta Limestone, Glendo Shale, Forelle Limestone, Difficulty Shale, Ervay, Freezeout Shale, and Little Medicine (fig. 4). The Goose Egg is unique in that most of its members were named before the formation was named. The Opeche and the Minnekahta were named as formations by Darton (1901) for exposures of the rocks in the Black Hills. Thomas (1934) described tongues of the western Wyoming Permian Phosphoria and Triassic Dinwoody Formations that extended eastward to central and southeastern Wyoming into the lower part of the Chugwater Formation. Thomas defined these, in ascending order, as the Sybille, Forelle, and Ervay Tongues of the Phosphoria and the Little Medicine Tongue of the Dinwoody; he also named the red beds between the

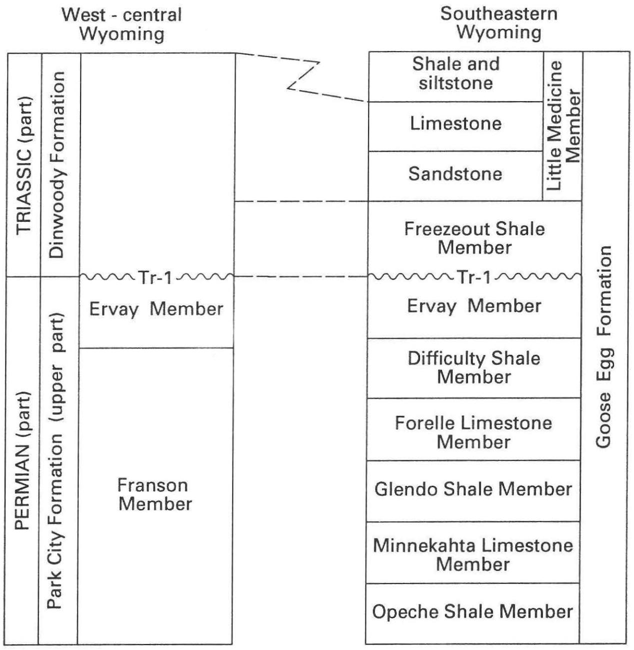


Figure 4. Generalized correlation of the Goose Egg Formation of southeastern Wyoming with the Park City and Dinwoody Formations of west-central Wyoming. Modified from Thomas (1934), Burk and Thomas (1956), Maughan (1964), and Boyd and Maughan (1973).

Forelle and the Little Medicine the Freezeout Tongue of the Chugwater. Condra and others (1940) extended the names Opeche and Minnekahta from the Black Hills to the Laramie Range. They also named the Glendo for deposits between the Minnekahta and the Forelle. When Burk and Thomas (1956) defined the Goose Egg, they abandoned the name Sybille (Thomas, 1934) for the lowest carbonate unit and used the name Minnekahta. As originally defined by Thomas (1934), the Freezeout was troublesome because where the Ervay was present the unit was divisible into two parts. To clarify this situation, Maughan (1964) named the lower part Difficulty and retained Freezeout for the upper part. In the southern Bighorn Mountains, Woodward (1957) and French (1959) reported a distinct carbonate, siltstone, and sandstone unit as present locally between the Pennsylvanian Tensleep Sandstone and the Opeche. This unit has been called the "Nowood member" of the Phosphoria (McCue, 1953; Tourtelot, 1953). The Permian-Triassic boundary, discussed later in this report, is thought to lie at, or near, the Ervay-Freezeout contact; thus, in this report special attention will be given to the top two members of the Goose Egg.

Within the Goose Egg, the amount of carbonate rocks increases to the west, and the amount of clastic rocks increases to the east; there also are more carbonate rocks in the Permian part of the formation than in the Triassic part. Within the individual marine tongues, there is an idealized, eastward-trending facies change from limestone, through dolomite, gypsum, and finally to halite (Maughan, 1964; Greer, 1985; Renner, 1988). The latter facies is reported for some tongues in the subsurface of the Powder River Basin and the Alliance Basin of Nebraska. Limestone in the Goose Egg is generally quite dolomitic and locally contains significant amounts of chert. Greer (1985) stated that, in the Laramie Range, gypsum is more abundant in thicker sections of the Goose Egg; in the subsurface, the sulfate mineral is anhydrite. The red-bed members of the Goose Egg are remarkably similar; in hand sample individual members cannot be distinguished (Renner, 1988). The deposits are composed mostly of brick-red, clayey siltstone or silty claystone and minor amounts of mudstone. Finer grained parts are darker red, and light-green reduction spots are common. The rocks are somewhat calcareous, and sedimentary structures are rare. The Goose Egg lies unconformably on the Pennsylvanian and Permian Hartville, Casper, and Minnelusa Formations in the Hartville Uplift, northern Laramie Range, and central Powder River Basin, respectively; in the Bighorn Mountains the Goose Egg lies unconformably on the Pennsylvanian Tensleep Sandstone. The contact between the Goose Egg and the overlying Red Peak Formation is generally considered conformable. Surface thicknesses of the Goose Egg vary considerably depending on how much gypsum has dissolved and on how much erosional relief is on the underlying unconformity. At the type locality the Goose Egg is 380 ft thick. In general, the Goose Egg thins northward along the eastern flank of the Bighorn Mountains. Some workers (Burk and Thomas, 1956; Faulkner, 1956; French, 1959) believed that this thinning is the result of lower members systemically dropping out by depositional onlap of onto Tensleep unconformity until only the upper part of the formation is present. This phenomenon, however, is not apparent on cross sections by Privrasky and others (1958).

The Goose Egg, or its lithostratigraphic equivalent, is present throughout eastern Wyoming, southeastern Montana, western South Dakota, and western Nebraska (Condra and others, 1940; Maughan, 1964). In the Black Hills, rocks equivalent to the Goose Egg are represented in the Opeche Shale, Minnekahta Limestone, and in the lower part of the Spearfish Formation. The individual members, as defined by Burk and Thomas (1956) and Maughan (1964), are generally identifiable in the Laramie Range and southern Bighorn Mountains. The members have not been adequately traced northward along the eastern flank of the Bighorn Mountains, and descriptions in the literature discuss the deposits only as the Goose Egg Formation or as a sequence of red shale and gypsum. Some of the lower members have been identified in the Hartville Uplift, but descriptions of rocks in the upper part of the Goose Egg sound more like descriptions of the lower part of the Spearfish in the Black Hills. The Opeche and Minnekahta are relatively easy to identify and trace in the subsurface of the southern and central Powder River Basin; however, because of the thinness of the marine tongues in the upper part of the Goose Egg and the scale of oil and gas logs individual members in this part of the formation are somewhat difficult to identify and trace in areas away from the southwestern corner of the basin. To the west, the Permian part of the Goose Egg correlates with the Park City, and the Triassic part of the Goose Egg correlates with the Dinwoody (fig. 4). In this respect, the Goose Egg may be thought of as a landward facies of these two marine formations. West of the Rattlesnake Hills, the rocks are more like the Park City and Dinwoody, and the name Goose Egg is not applied. No fossils have been reported from the red-bed members of the Goose Egg, and only undiagnostic fossils have been reported from some of the carbonate members. The age of the formation has been determined by correlating its carbonate members westward into the Park City and Dinwoody Formations, the ages of which are known. Maughan (1980) placed the Opeche and Minnekahta in the Leonardian and the higher Permian members in the Guadalupian; the Freezeout and Little Medicine were assigned to the Early Triassic.

The eastward regional thinning and eastward carbonate to halite mineral progression of the nonclastic members of the Goose Egg suggest that marine invasions from the west were the active factors in the intertonguing rather than increased progradation of clastic detritus from the east. The great areal extent of the nonclastic units implies that the invasions occurred rapidly and that the transgressed surface was one of very low relief; the thinness of the units indicates either that the invasions were of relatively short duration or that sediment accumulation rates were very low. Each of these members probably represents one regional transgressive-regressive cycle of marine water from the miogeocline (Sheldon, 1963; Peterson, 1980; Greer, 1985), although minor sea-level fluctuations probably occurred during deposition (Rath, 1981). The bulk of each unit probably formed during the transgressive phase; however, the gypsum that is observed locally on the top of the Minnekahta Limestone in the Black Hills (Braddock, 1963) probably represents deposition during regressive conditions. The lack of a clastic fraction and the very thin laminations of the carbonate rocks required clear water and low-energy conditions, respectively. Moreover, the chemical precipitation of carbonate and evaporite minerals required shallow water, with little or no inflow of fresh water and a warm and arid climate in order to form hypersaline water. These rocks were probably formed in a shallow-marine environment in intertidal (tidal flats) and subtidal (lagoons) settings. The observed mineral progression is probably due to increased salinity associated with higher evaporation in the shallower parts of the environment in the landward direction. In addition, some geographic isolation from the open-marine conditions to the west was probably necessary. This could have taken the form of restricted water bodies along a wide, transitional coast line. Greer (1985) even suggested a topographic sill in central Wyoming to facilitate these isolated conditions. Local differences in mineralogy could be due to small-scale variations in bathymetry or currents. A shallow-marine environment is favored over a supratidal or sabkha setting because features associated with these environments are lacking (Watson, 1983; Greer, 1985).

The red-bed members of the Goose Egg probably accumulated under depositional conditions similar to those of the overlying Red Peak Formation red beds. In this respect, the Goose Egg red beds represent the beginning of the Red Peak depositional setting that resulted in the great mass of red beds in central and eastern Wyoming. The debate over the specific depositional environment of these red beds has continued for years, and the topic will be discussed in more detail later in this report. The surface upon which the Goose Egg red beds were deposited must have been very flat, gently sloping to the west, and directly adjacent to a marine shelf on the west. The fine grain size and general lack of sedimentary structures indicates low-energy conditions, and the lack of any paleosols or body and trace fossils might indicate an environment hostile to life. Some workers have advocated a shallow-marine environment for these deposits (such as Privrasky and others, 1958); however, the deposits are probably nonmarine. Their position between marine tongues of the miogeocline, some showing transgressive-regressive cycles, supports this conclusion. Renner (1988), noting the predominance of silt-sized detritus, lack of fluvial-channel deposits, lack of current ripples, and absence of any evidence of life, suggested that perhaps these red beds were, at least in part, eolian. The source area for most of the detritus in the Goose Egg red beds was probably the exposed craton to the east, although some material might have been sourced to the south in the Ancestral Rocky Mountains. The cause of the cyclic marine invasions that influenced Goose Egg deposition is unknown. Similar cycles are present in certain Permian deposits in the Permian Basin of western Texas and eastern New Mexico (Meissner, 1967); perhaps the phenomenon is at least regional rather than local. Further studies might reveal that the cycles are, in fact, global in scale and related to interglacial stages of Gondwana glaciation.

Maughan (1964) restricted the Freezeout Shale Member to those red beds between the Ervay and Little Medicine Members as exposed in the Freezeout Hills in northeastern Carbon County, Wyoming. Lithic properties in the Freezeout are horizontally and vertically homogeneous. The member is best described as a brick-red, fissile to blocky siltstone, with some very finely laminated silty claystones. In addition, Privrasky and others (1958) mentioned that the unit can locally be dolomitic or gypsiferous. The contact with the underlying Ervay is sharp and appears conformable but is apparently disconformable. The Freezeout is 10–15 ft thick in southeastern Wyoming (Maughan, 1964); it is 35 ft thick at its type section and 48 ft thick at the Goose Egg type section. The member is reported by French (1959) to be 50–60 ft thick in the southern Bighorn Mountains. Woodward (1957) reported a subsurface thickness of the member in the southwestern Powder River Basin of 55 ft. The Freezeout is easily traced westward in the subsurface. The member correlates with yellowish calcareous siltstone in the lower part of the Dinwoody Formation in central Wyoming (Maughan, 1964).

The Little Medicine Member was defined by Thomas (1934) for exposures at Flat Top anticline, 8 mi north of Medicine Bow, Wyoming. At the Goose Egg type locality, Burk and Thomas (1956) noted that the Little Medicine consists of three parts.

1. Upper part—Red, sandy shale and siltstone interbedded with minor amounts of very fine grained sandstone and white gypsum (10 ft)
2. Middle part—Mottled green to purple or gray, finely crystalline, sandy, rippled limestone interbedded with minor amounts of green shale (10 ft)
3. Lower part—Green to tan, fine to very fine grained, rippled sandstone interbedded with minor amounts of green shale (8 ft)

The Little Medicine is known for its variegated color of carbonate rocks and evaporites and relatively high content of mica flakes in the clastic fraction (Thomas, 1934; Woodward, 1957). Woodward (1957) remarked that the high amount of mica in the member might indicate that metamorphic rocks were exposed in the source area. The upper part of the member is commonly rippled, and glauconite is present locally. In the southernmost part of the Bighorn Mountains, the member is reported by Woodward (1957) to thin eastward and change from a pink, sandy, dolomitic limestone to red, calcareous siltstone with interbeds of massive gypsum. Gypsum increases northward along the Red Wall, which is along the southern part of the eastern flank of the Bighorn Mountains, and McLellan (1968) mentioned some limestone in the member in this area. The contact of the Little Medicine with the underlying Freezeout is gradational and conformable; the contact with the overlying Red Peak Formation is sharp and generally considered conformable. Because the Red Peak represents a westward-prograding deposit that followed the retreat of the Little Medicine sea, this contact probably climbs stratigraphically and becomes younger toward the west. In the southern Bighorn Mountains, McLellan (1968) used the top of the highest limestone or gypsum in the Little Medicine as the contact. I would modify this to the highest significant limestone or gypsum bed because where I have observed this contact minor amounts of gypsum are commonly present in the basal part of the Red Peak. Should the Little Medicine be missing because of nondeposition, picking the Goose Egg–Red Peak contact might be difficult, especially in the subsurface. Both the

Red Peak and the Freezeout contain horizontally bedded red siltstone, and gypsum can be present on either side of the contact. Because of these problems, one might accidentally pick the contact at the top of the Ervay (or, if this is also missing, on an even lower unit) and, hence, mistakenly place the contact too low. In southeastern Wyoming, the Little Medicine is 5–25 ft thick. In the southernmost part of the Bighorn Mountains, Woodward (1957) reported the member to vary between 5 and 20 ft thick in surface sections; a subsurface measurement revealed 15 ft of the member. The Little Medicine correlates with the upper part of the Dinwoody of west-central Wyoming (fig. 4). The sandstone and limestone in the lower and middle parts of the Little Medicine of Burk and Thomas (1956) represent the Little Medicine Tongue of the Dinwoody Formation as defined by Thomas (1934) in central Wyoming; the red argillaceous rocks in the upper part of the Little Medicine might better be placed in the base of the Red Peak. Southeast of the Little Medicine type section in the central and southern Laramie Range, the member is mostly gypsum. Greer (1985) has traced the member in the subsurface south into the northern Colorado Front Range, and E.K. Maughan (U.S. Geological Survey, retired, written commun., 1992) stated that south of the Colorado-Wyoming State line the unit grades into a thin siltstone bed that might extend as far south as Lyons, Colorado. In the southern Bighorn Mountains, the Little Medicine is widespread but is usually the poorest developed member of the Goose Egg. French (1959) identified the unit in the subsurface just north of Kaycee, Wyoming. The farthest north that I observed the member to crop out in the Bighorn Mountains was in the southern Red Wall area about 2 mi northeast of Deadman Butte (fig. 5), but the member is reported to extend farther north (Privrasky and others, 1959; McLellan, 1968). Maughan stated that the member grades into a thin siltstone bed in the northern Bighorn Mountains that can possibly be traced as far north as Yellowtail, Montana. Maughan also remarked that the member can be easily traced in the subsurface of the northeastern Powder River Basin where it is composed of gypsum and, possibly, some halite.

Contact of the Permian and Triassic Systems

An almost worldwide hiatus separates the Paleozoic and Mesozoic Erathems. This gap in the stratigraphic record is well supported by a significant faunal discontinuity, but physical evidence of a disruption is meager or lacking in some areas. In most of the Western Interior, however, this unconformity is characterized by considerable erosional relief. In this region the unconformity, the Tr–1 of Pipiringos and O'Sullivan (1978), usually separates Permian rocks of Guadalupian age from rocks of Early Triassic age. Most of the missing time is probably Permian because rocks representing the youngest stage of the Permian, the Ochoan, are never recognized. Pipiringos and O'Sullivan (1978) stated that the time necessary to create this break in the stratigraphic record could have been as little as 1–2 million years or as much as 5–6 million years. In eastern and central Wyoming, no physical or faunal evidence has been reported to support this unconformity, and the existence of the surface is based on evidence extrapolated from outcrops to the west. In west-central Wyoming physical evidence of an unconformity between Permian and Triassic rocks is subtle, but faunal evidence exists. The highest carbonate bed in the Ervay Member of the Park City Formation contains late Guadalupian-age (early Capitanian) conodonts (Wardlaw and Collinson, 1986), and the lowest part of the Dinwoody Formation contains early Scythian-age (Griesbachian) conodonts (Paull and Paull, 1983). Thus, in the southwestern Powder River Basin, the unconformity should lie between the Ervay and Freezeout Members of the Goose Egg Formation. A few workers doubt the existence of an unconformity in eastern Wyoming at this or any level because of the lack of physical evidence and call for continuous deposition across the Permian-Triassic boundary. For example, Renner (1988) concluded that the boundary between these two systems is conformable, and he presented a depositional model to support his idea. Renner also noted that the upper part of the Ervay Member of the Park City in west-central Wyoming reportedly records westward regression, which would indicate that the top of the Ervay Member of the Goose Egg in eastern Wyoming might be slightly older that its counterpart to the west. If this is the case, the Permian-Triassic boundary in eastern Wyoming might lie somewhere in the lowest part of the overlying Freezeout Member. Based on regional evidence, however, most workers believe that a hiatus at the close of the Permian created a disconformity that exists undetected at or near the Ervay-Freezeout contact, and the term paraconformity is commonly applied. This assignment of the level of the Permian-Triassic contact dates back to Thomas (1934). Maughan (1964) mentioned that a southward thinning of the Ervay Member of the Goose Egg in the Laramie Basin might be erosional rather than depositional and thus might represent a slight regional angular unconformity at this level. In addition, E.K. Maughan (written commun., 1992) stated that in the eastern Owl Creek Mountains of central Wyoming a silcrete is present on the upper surface of the Ervay that he believes formed during the postulated hiatus. Because accurately placing the Permian-Triassic contact depends on correctly identifying the Ervay, a problem commonly exists in areas beyond the southwestern part of the Powder River Basin where physical correlation of the members of the Goose Egg back to the Goose Egg type section is lacking. Moreover, should the Ervay be missing because of nondeposition, the Triassic Freezeout Member would be in contact with the Permian Difficulty Member,

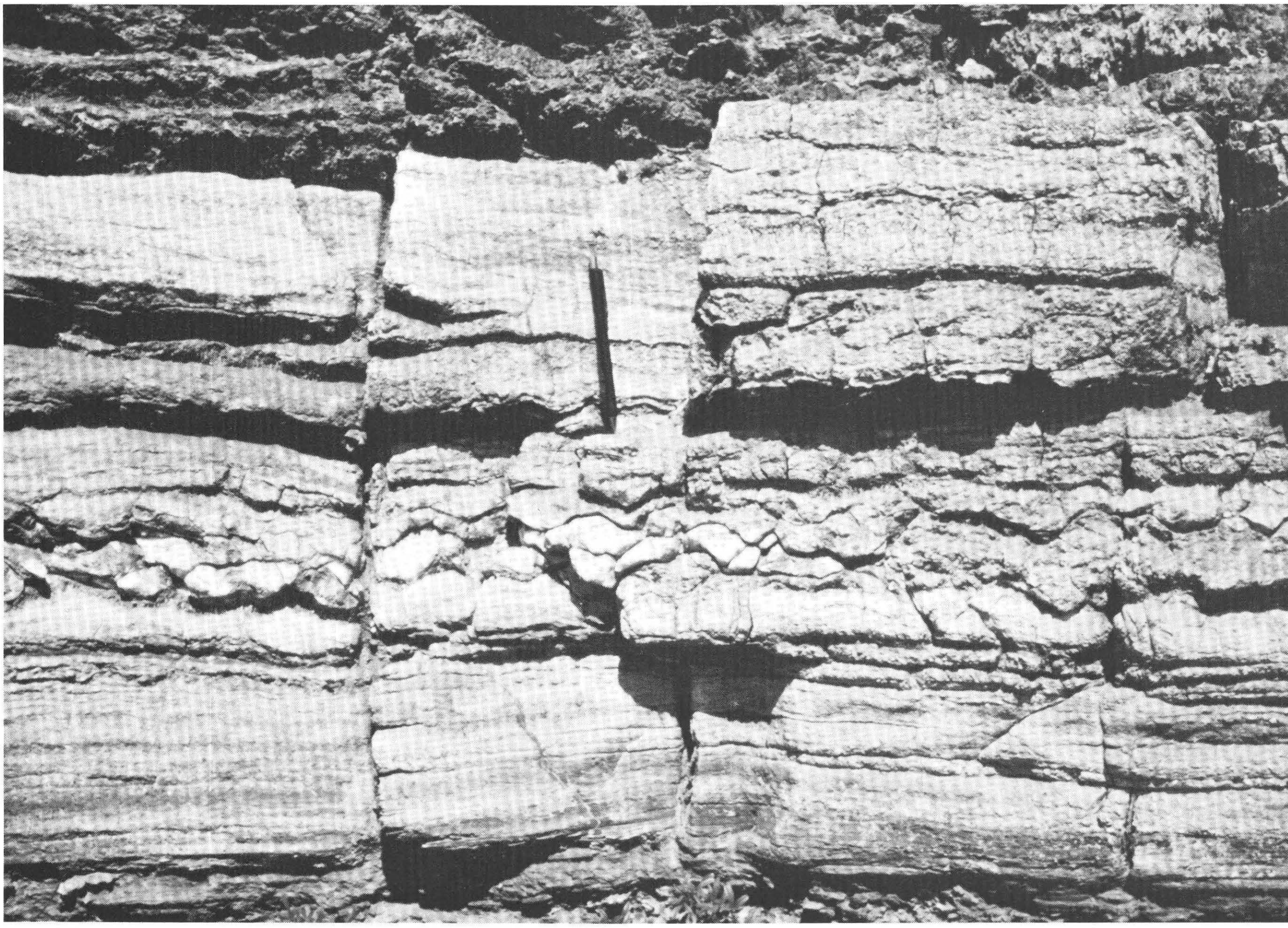


Figure 5. Gypsum bed in the Little Medicine Member of the Goose Egg Formation, 4 mi northeast of Deadman Butte, southeastern Bighorn Mountains. Pencil shown for scale.

and this juxtaposition of parallel-bedded, red siltstone would make correct placement of the contact very difficult. Burk and Thomas (1956) mentioned that argillaceous rocks in the lower part of the Red Peak Formation tend to be lighter red, slightly less sandy, less resistant, and less gypsiferous than similar rocks in the upper part of the Goose Egg. In addition, these workers reported that on geophysical logs argillaceous rocks in the Red Peak have a much lower resistivity and a higher natural gamma ray than similar rocks in the Goose Egg. Correctly locating the Permian-Triassic contact within the Spearfish in the Black Hills is particularly troublesome and will be discussed later in this report.

Early Triassic Progradation

During the Early Triassic, a large volume of mainly silt sized sediments prograded westward across eastern and central Wyoming. This event dominated the history of Lower Triassic rocks and resulted in the spectacular exposures of red rocks now visible in the area.

Paralic Silty Rocks

Triassic red beds are exposed in the northern Laramie Range and Hartville Uplift, along the eastern flank of the Bighorn Mountains, and around the flanks of the Black Hills. On the south and west sides of the Powder River Basin the greatest volume of the red beds makes up almost all of the Red Peak Formation in the lower part of the Chugwater Group. Love (1939) named this formation for exposures near Red Peak on the northwestern margin of the Wind River Basin.

In the area of the Powder River Basin, the red beds consist mostly of siltstone and silty claystone in alternating beds of differing induration (fig. 6); mudstone, sandstone, and shale are less abundant. In general, the unit coarsens upward, and some individual subdivisions of the unit also coarsen upward. The rocks are carbonate cemented, and some discontinuous lenses of gypsum present locally near the base of the unit probably represent the final phase of Little Medicine regression. Well-rounded mica grains are common on bedding surfaces, and illite is the dominant clay mineral. The color of the rocks is best described as brick red, and finer grained rocks are generally somewhat darker red. Greenish-gray reduction spots and bedding-parallel bands are common in the claystones; McLellan (1968) reported that some bands could be traced for as much as a mile. In coarser grained rocks, reduction discoloration commonly is present parallel with joints and bedding planes. The red color is from hematite staining that exists as grain coatings and in interstitial spaces. Current thinking (Reading, 1986) is that the original sediments were not red but contained a significant amount of iron-rich minerals such as magnetite, ilmenite, biotite, and hornblende. Following deposition in a warm and wet environment, the iron-rich minerals broke down into clays and immature iron oxides and hydroxides, which then percolated downward and coated the fragments; these minerals later converted to hematite. Thus, the color is authigenic, and it developed in situ during an early stage of diagenesis. Bedding in the red beds is parallel, and some individual beds can be traced for long distances. Sedimentary structures include parallel to ripple lamination, convolute lamination, small-scale trough cross-stratification, and small-scale cut and fill features. Bed forms include subtly asymmetrical ripple marks and mud cracks. Biogenic features include bioturbated layers and rare, isolated vertical burrows.

In central Wyoming, Picard (1967) divided the Red Peak into the following five facies in ascending order: silty claystone facies, lower platy facies, alternating facies, upper platy facies, and sandy facies (see also Picard, 1964; Picard and Wellman, 1965). High and Picard (1967) later removed the sandy facies from the Red Peak and placed it in the bottom of the Crow Mountain Formation as defined by them. Removing the sandy facies from the Red Peak does conveniently restrict the Red Peak to containing only red beds. The best exposures of the red beds are on the west side of the Powder River Basin. A study of the Red Peak was conducted by McLellan (1968), a student of Picard's, as part of a comprehensive investigation of the Chugwater Group along the east flank of the Bighorn Mountains from Arminto, Wyoming, to Yellowtail, Montana. McLellan used the four facies divisions of High and Picard (1967) in his study and found that the facies were consistent and traceable along the entire length of the Bighorn Mountains. Considering the known regional uniformity of the red beds, and the fact that this slice of the unit is probably subparallel to depositional strike, this continuity is not unexpected. A summary of McLellan's facies descriptions is as follows.

1. Silty claystone facies—Silty claystone, especially in the lower part, with siltstone increasing upward; some claystone; very rare sandstone and intraformational claystone conglomerate; thin, parallel bedded; burrowed; weathers platy to blocky; silty claystone is parallel to wavy laminated with some disturbed bedding and mud cracks; siltstone is poorly sorted, cross-stratified, and ripple laminated, with ripple marks on bedding surfaces; contact with the Goose Egg is sharp; 105–131 ft thick; facies characteristics are laterally persistent (this is the least indurated facies of the red beds and usually forms a grass-covered strike valley known in this area as the red valley)

2. Lower platy facies—Siltstone, some silty claystone, and rare sandstone; silty claystone decreases upward and sandstone increases; parallel to wavy bedded; cross-stratified; burrowed; ripple marks on bedding surfaces; rare mud cracks; some intraformational claystone conglomerate; siltstone is poorly sorted, and weathers platy to flaggy; contact with the underlying facies placed where siltstone exceeds silty claystone; 140–280 ft thick; facies characteristics are laterally persistent (this facies is more resistant that the underlying facies and generally forms the lowest exposures on the slopes out of the strike valleys)

3. Alternating facies—Siltstone, some sandstone, and rare claystone; alternating massive, very thick to thick bedded siltstone, and platy to flaggy siltstone intervals; siltstone becomes sandier upwards and sandstone increases; parallel bedded; burrowed; massive siltstone is well sorted, parallel laminated or cross-stratified and makes up 64–80 percent of the facies; platy to flaggy siltstone is usually poorly sorted, wavy laminated, cross-stratified, and ripple marks are present on bedding surfaces; contact with underlying facies placed at the base of the lowest significant massive siltstone bed; 205–338 ft thick; facies characteristics change laterally (this facies forms the conspicuous cliffy part of the valley wall)

4. Upper platy facies—Siltstone, claystone, and minor amounts of sandstone and intraformational claystone conglomerate; micaceous; parallel to wavy bedded; cross-stratified; burrowed; siltstone is poorly sorted, shows ripple marks on bedding surfaces, and weathers platy, flaggy, or slabby; contact with the underlying facies placed at the base of the lowest significant sequence of platy siltstone; 8–19 ft thick; facies characteristics are laterally persistent (this facies commonly forms the wall above the ledge capping the cliffs in areas overlain by the resistant Alcova Limestone)

McLellan also stated that sandstone in the red beds is arkosic, very fine grained, subangular to subrounded, and silty; siltstone is feldspathic to arkosic and subangular; claystone weathers platy to blocky; and all of the rock types are cemented with carbonate. In addition, McLellan reported that most ripple marks are asymmetrical, that cross-stratification occurs as small-scale troughs, and that convolute laminations are common. It should be noted that the description of Cavaroc and Flores (1991) of the Red Peak along a shorter segment of the unit in the southeastern Bighorn Mountains mentions more sandy rocks and high-energy sedimentary structures than reported by McLellan. Oil-stained rocks have been observed in the lower part of the Red Peak in the Bighorn Basin (Picard,

Figure 6. Alternating beds of resistant siltstone and less resistant silty claystone in the Red Peak Formation, southeastern Bighorn Mountains. Lowest ledge in foreground about 6 ft high.

1975), and oil shows are reported from the interval in some test wells (Picard, 1978).

In the Black Hills, rocks equivalent to the red beds of the western Powder River Basin are contained in the upper part of the Spearfish Formation. The facies divisions of High and Picard (1967) have not been applied in the Black Hills. The Spearfish has been divided by numerous workers (Reeside and others, 1957; McKee and others, 1959; Mapel and Pillmore, 1963; Wolcott, 1967; Babcock, 1967), and the following diminutive scheme is based on an integration of their efforts.

1. Lower mudstone facies—Red mudstone; secondary gypsum; massive; reported average thickness 78 ft in the southern part of the Black Hills and 81 ft in the northern part; green shale or carbonaceous shale sometimes described at the top of the unit in the southern part of the Black Hills (the lower mudstone facies is probably equivalent to the Glendo Shale Member of the Goose Egg)
2. Evaporite facies—Massive, resistant, white gypsum beds separated by intervals of red mudstone; minor dolomite and limestone (some stromatolitic) noted; reported average thickness 174 ft in the southern part of the Black Hills and 58 ft in the northern part (unit thins by decrease in amount of clastic rocks); individual gypsum beds average about 18 ft thick (all of the bedded gypsum in the Spearfish is included in this facies; probably equivalent to that part of the Goose Egg that lies above the Glendo Shale Member)
3. Upper mudstone facies—Red mudstone; minor sandstone; facies coarsens upward; secondary gypsum; in the northern part of the Black Hills the lower part of the facies is massive and the upper part contains some cross-stratification; there is a reported thickness of 90 ft in the southern part of the Black Hills and 155 ft in the northern part (this facies is probably equivalent to most, if not all, of the red beds in the Red Peak in the Bighorn Mountains)
4. Sandy facies—Red, very fine grained sandstone; some mudstone; cross-stratification and ripple lamination noted; reported thicknesses 80–200 ft, but the top of the facies is always beveled by erosion (this facies might be equivalent to the upper part of the red beds in the Bighorn Mountains, but just as likely it might correlate with the sandy rocks at the top of the Red Peak)

Specific sedimentary features reported in the Spearfish include mud cracks, external molds of salt crystals, intraformational claystone conglomerate, and rare, small-scale, sandstone channel bodies. In general, the Spearfish is less resistant than the Red Peak, and all but the uppermost part of the formation tends to form a strike valley resulting in the topographic feature known as the race track that encircles the Black Hills. The top of the Spearfish is everywhere truncated by Jurassic erosion. The Alcova Limestone and stratigraphically higher formations of the Chugwater Group have not been identified in the Black Hills; these units either never extended as far east as the Black Hills, or they have been removed by erosion. Four significant gypsum beds are usually reported in the evaporite facies in the southern part of the Black Hills; in the northern part, one to three beds are reported. There seems little doubt that these beds are related to the marine tongues of the Goose Egg that exist in the southern and western parts of the Powder River Basin; however, no published correlations of the marine members of the Goose Egg with the evaporite units of the Spearfish exist. Suffice it to say that the Permian-Triassic contact probably lies somewhere in the upper part of the evaporite facies or in the lower part of the upper mudstone facies.

With the exception of rare trace fossils and bioturbation, almost no evidence of plant or animal life exists in the red beds of central and eastern Wyoming. Love (1948) reported the finding of one poorly preserved bivalve in the red beds on the northern margin of the Wind River Basin. A few footprints of reptiles or amphibians were described by Lull (1942) at a locality near the northeastern margin of the Wind River Basin, but their identity is disputed (Branson, 1947; Colbert, 1957). Fish scales have been reported near Thermopolis, Wyoming (Picard, 1967), and burrows have been described on the east flank of the Bighorn Mountains (McLellan, 1968). The lack of body fossils might be because of an environment of deposition that was hostile to life. Perhaps high-salinity conditions limited the number of plant and animal species that could exist or maybe the physical environment was so active, with clastic fragments being reworked on a daily bases, that an adequate food chain never developed. However, it is hard to visualize such a broad depositional surface, existing for millions of years under humid climatic conditions, that would not be thriving with life. More likely it was oxidizing conditions that prevented preservation.

The contact between the red beds of the Red Peak and the Goose Egg Formation is sharp and generally considered conformable, although McLellan (1968) thought it was disconformable along the east side of the Bighorn Mountains. Correctly, the contact should be placed on the top of the Little Medicine Member of the Goose Egg, but in areas where this unit cannot be correctly identified or is missing because of nondeposition the contact is, by necessity, placed on the top of the highest carbonate or evaporite unit in the Goose Egg. In the Spearfish of the Black Hills, no unusual contact has been described in the evaporite or upper mudstone facies, and it is assumed that rocks equivalent to the red beds of the Red Peak on the west side of the Powder River Basin rest conformable on older strata of the Spearfish. On the west side of the basin, the contact of the red beds with the overlying sandy unit, which is traditionally also placed in the Red Peak, is sharp and conformable. This is also true of the contact between the upper mudstone and sandy facies of the Spearfish. On geophysical logs, the main body of the red beds is relatively easy to identify. The lower boundary is usually placed at the top of the highest high-resistivity kick below the red beds, and the upper boundary is placed below the high-resistivity kick of the Alcova Limestone. In areas where the Alcova is missing because of nondeposition or erosion, placing this contact can be problematic.

In the area of the Powder River Basin, the upper part of the red beds is commonly beveled by Jurassic erosion, and published thicknesses might be suspect. Complete sections of the red beds average about 615 ft thick in the southern and central segments of the eastern flank of the Bighorn Mountains, and to the north in the Montana segment of the flank a measurement of 641 ft is recorded (McLellan, 1968). In the northern Laramie Range, reported complete thicknesses of the red beds average 635 ft (Sears, 1949; Jenkins, 1950; Schwarberg, 1959). The red beds thin toward the east, and in the Black Hills, equivalent rocks, the upper mudstone facies of the Spearfish, are less than 160 ft thick (Braddock, 1963; Babcock, 1967). West of the Bighorn Mountains, the red beds thicken to as much as 1,100 ft in west-central Wyoming (Picard, 1967, 1978). According to Picard and others (1969), the red beds are equivalent to the Woodside Formation and most of the overlying Thaynes Formation in western Wyoming (fig. 7).

Individual units within the red beds can be traced on the surface over long distances. Picard and Wellman (1965) correlated one 5.5-ft-thick bed of claystone more than 73 mi in an east-west direction, and they estimated that the bed extends for more than 110 mi in the north-south direction. In the subsurface, the red beds can be correlated over even greater distances by the recognition of several groups of high-resistivity spikes on geophysical logs. Two marker beds were correlated by Renner (1988) in the subsurface for a distance of 280 mi. J.D. Love (U.S. Geological Survey, retired, oral commun., 1990) stated that one particular 3-ft-thick bed can be identified over a 50,000 mi^2 area. Corresponding spikes are common on the natural gamma ray trace, and Renner (1988), using a hand-held scintillometer to construct gamma-ray profiles, identified some of these beds on the surface. From this, he determined that these spikes are due to an abnormally low clay content in the marker beds. Lacking any diagnostic fossils in the red beds themselves, the age of these rocks must be inferred from regional correlations. Because the red beds correlate with the Woodside and most of the Thaynes Formation (fig. 7) and because these formations are Early Triassic, it is assumed that this is also the age of the red beds in central and eastern Wyoming.

Published paleocurrent measurements are few, and they were collected mostly from the eastern flank of the Bighorn Mountains. Some of these measurements were taken from trough cross-strata, and, because the reliability of the measuring technique used cannot be assured and because the meaning of these values in a paralic environment is controversial, I attach little significance to these data. Measurements from ripple marks, however, might provide information on the general trend of the shoreline if they were formed by waves or onshore currents. A predominance of ripple-train measurements with long axes trending northeast might indicate a shoreline of the same orientation.

The depositional conditions under which the red beds were deposited has been debated for years, and many ideas have been put forward. Postulated environments include offshore marine shelf, delta front, tidal flat, coastal plain, flood plain, and others. There are, however, various depositional facts that are generally accepted as true, and they are considered basic to the understanding of the depositional history of these deposits. Clastic progradation probably occurred from an eastern source area toward a marine environment to the west. Depositional energy probably increased slightly toward the east. McLellan (1968) remarked that intraformational claystone conglomerate in the red beds, which he attributed to the reworking of mudcrack clasts by current action, increases eastward in the Bighorn Mountains. In addition, what little sandstone there is in the red beds increases eastward (Picard, 1966; McLellan, 1968). The surface upon which these sediments accumulated must have been very broad, almost flat, and probably just above or just below sea level. Tectonically, the region was relatively stable, but subsidence of the surface must have occurred for such a great thickness of sediment to have accumulated. Both subaqueous and subaerial structures are present in the rocks. Ripple marks formed in shallow water, and mud cracks indicate that the surface was episodically exposed above water. Any theory on the formation of these red beds must explain the regional nature of the depositional environment that resulted in the persistent, widespread, lithogenetic units that are so typical of these deposits.

The salient question has always been: are these deposits marine or nonmarine? Traditionally, red beds have been considered nonmarine, yet most workers have assigned marginal-marine environments to the Triassic red beds of central and eastern Wyoming. The generally coarsening upward nature of the red beds does support a marginal-marine setting, and, because three of the four facies described by High and Picard (1967) also coarsen upward, fluctuations of the shoreline probably occurred. Picard (1967) assigned a paralic environment to his silty claystone facies, a tidal flat to his lower platy facies, a nearshore marine to his alternating facies, and a tidal flat to his upper platy facies. Cavaroc and Flores (1991) assigned a distal delta front environment to the lower part of the Red Peak, delta front for a higher part, and fluvial (including the channel-fill deposits of sandy braided streams) for the upper part. Unfortunately, the red beds of central and eastern Wyoming do not exactly match any described marine or nonmarine deposit.

I favor an expanded version of a depositional setting first considered by Burk (1953) that calls for several

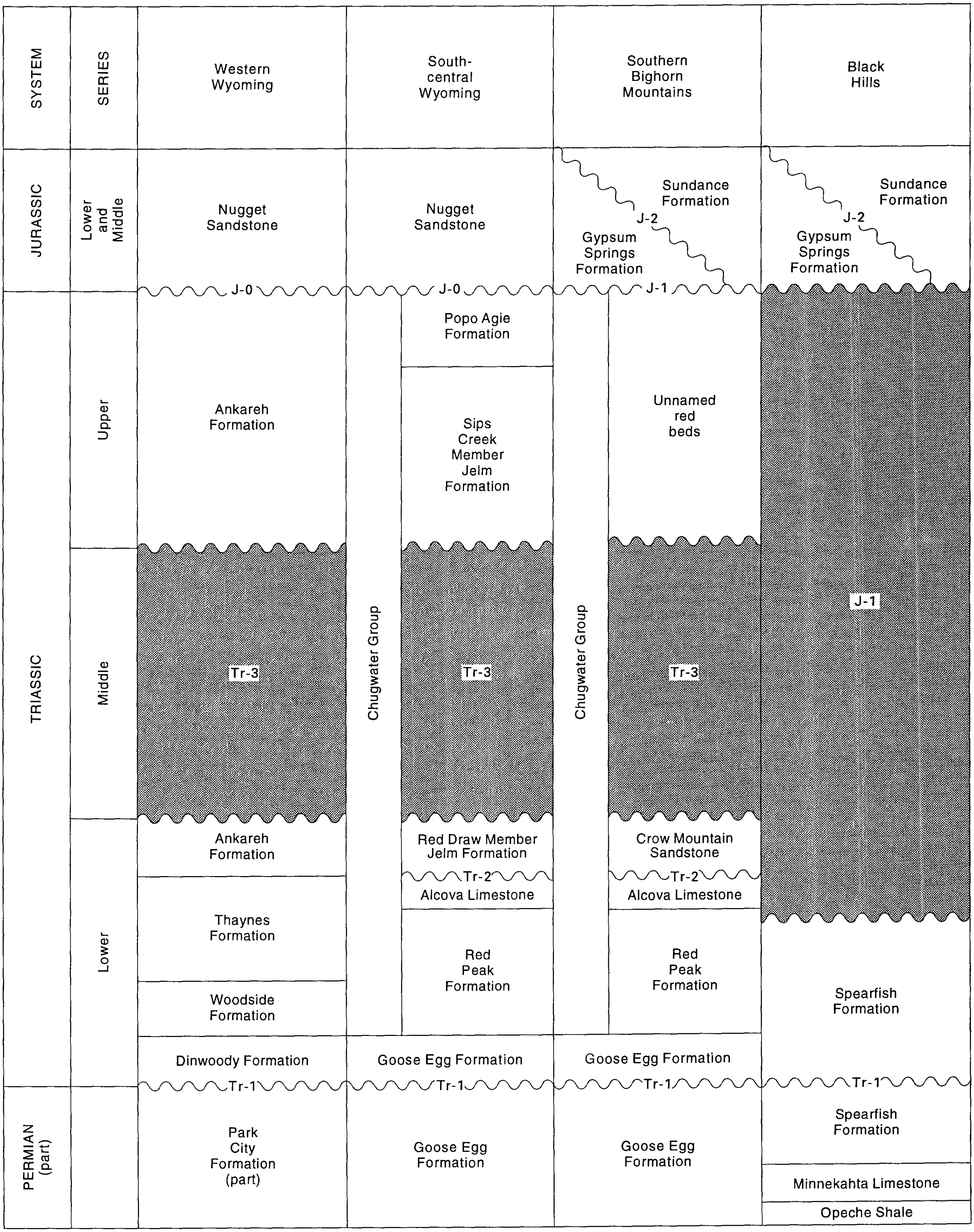


Figure 7. Basic correlation of Triassic units across Wyoming. Tr–1, Tr–2, Tr–3, J–0, J–1, and J–2 represent unconformities. Modified from Pipiringos (1968), Boyd and Maughan (1973), Pipiringos and O'Sullivan (1978), and Carr and Paull (1983).

processes operating simultaneously on different parts of the depositional surface. The eastern part of the surface was at, or just above, base level and was influenced by low-energy fluvial processes in the form of very broad, shallow, sheet floods that continuously changed position, thereby gently reworking the sediments and slowly moving the material to the west. The central, and largest, part of the surface took the form of a giant silt flat that was influenced by low-energy tidal action as each tidal cycle moved a thin veneer of water a great distance back and forth across this low topographic surface, slowly reworking the sediments with each pass. The western part of the surface was probably a shallow marine shelf. The shoreline probably trended northeast. Processes acting on the eastern part of the surface would explain the rarity of riverine-type fluvial features in the red beds, such as channel bodies with scour bases, and would provide a method of delivering a uniform amount of a given sediment type along an extended segment of the coast line as a sheet of detritus rather than as detritus dispersed from a channelized point source. Tidal processes acting on the central part of the surface would account for the observed ripples. Additionally, marine processes acting on the western part of the surface might have further aided in the distribution of detrital increments that ultimately resulted in the widespread continuity of individual rock packages. Subtle transgressive-regressive cycles might have shifted this entire regional setting back and forth, further complicating the eventual deposit. It is probable that no modern analog for this depositional setting exists today, at least not of the scale implied by these deposits. A remaining question is: if the red beds represent fine-grained deposition on an almost flat coastal plain and adjacent marine shelf, then where to the east are the coarser grained equivalents of these rocks in areas more proximal to the source area?

The source area for the sediments in the red beds is also an enigma. Most workers believe that the major source area was the exposed craton to the east. Some sediments might have come from the Ancestral Rocky Mountains to the south or possibly from other smaller areas to the north. A distant source area and long transport history would explain the predominance of fine-grained detritus in the red beds; however, the high feldspar and mica content noted in the red beds does not indicate a mineralogical maturity associated with a long transport history under humid climatic conditions. A closer source area is possible if the terrain was of low relief and the source material composed of fine-grained sedimentary rocks containing relatively fresh feldspar and mica. Red beds in the Permian part of the Goose Egg Formation might fit this requirement, and Nelson (1952) noted that the heavy minerals in the red beds of the Goose Egg and Red Peak Formations are similar. The problem with this idea is that the volume of red beds in the Permian part of the Goose Egg is quite small in comparison to that of the Triassic red beds, and it is doubtful that feldspar and mica eroded from the Permian Goose Egg could have survived the rigor of a second transport history, especially considering the hiatus that separates the Permian and Triassic red beds. And, the similarity in heavy minerals between the Permian and Triassic red beds probably just indicates a common source area. The feldspar and mica in the Triassic red beds might indicate that at least some acidic igneous or crystalline metamorphic rocks were exposed in the source area. Two problems with a feldspar- and mica-rich source area supplying detritus under humid climatic conditions are the apparent lack of kaolinite (a product of the leaching of acidic rocks) and clay-sized sediment (the product of soil-forming processes) in the red beds. It is possible, however, that the clay-size fraction bypassed the area that is now eastern and central Wyoming and was deposited along with carbonate minerals in the Dinwoody and Thaynes Formations. The high feldspar and mica content also might indicate physical weathering in advance of chemical weathering followed by rapid erosion. This process probably would require some amount of topographic uplift in the source area; however, the small grain size of the red beds does not support such tectonism.

Nearshore Marine Sandy Rocks

Throughout eastern and central Wyoming a sandy facies commonly is present at the top of the Red Peak Formation between the silty red beds of the Red Peak and the overlying Alcova Limestone (fig. 8). In the area of the Powder River Basin, the facies is well represented along the eastern flank of the Bighorn Mountains and in the northern Laramie Range. The facies is referred to in this report as the sub-Alcova sandstone. The rocks consist of very fine grained sandstone (subarkose) and silty sandstone, and the unit commonly coarsens upward. In comparison to rocks lower in the Red Peak, these rocks are calcareous, coarser grained, better sorted, and less micaceous; individual grains are better rounded. The rocks are commonly variegated in light shades of red, purple, and green. Color banding is common and commonly cuts obliquely across bedding at high angles. Downward-percolating fluids passing through the overlying limestone are probably responsible for the coloration. The rocks usually appear structureless or faintly convolutely laminated or turbated. Locally, the unit is parallel bedded or cross-stratified. In most places the unit forms a small vertical cliff tucked under the resistant Alcova. The contact of the unit with the underlying silty red beds is usually gradational and is conformable. The contact with the overlying Alcova is sharp and probably conformable. In central Wyoming the sub-Alcova sandstone is 4–15 ft thick (Picard, 1967). It is 3–11 ft thick in the northern Laramie Range (Sears, 1949; Schwarberg, 1959). Along the eastern flank of the Bighorn Mountains, the facies is as thick as 30 ft (Mapel, 1959), but it is usually about 10 ft thick, and it can be missing altogether (McLellan, 1968). Picard and others (1969) correlated this facies with part of the sandstone and limestone unit at the top of the Thaynes Formation in western Wyoming. When Picard (1964) first subdivided the Red Peak, he included the sub-Alcova sandstone at the top of the formation (variegated zone). Later, High and Picard (1967) moved this interval into the base of the Crow Mountain Sandstone (variegated sandy facies) in their revised definition of the formation (this reorganization of the Crow Mountain is not used by the U.S. Geological Survey). High and Picard reasoned that the facies is more closely associated with the Crow Mountain than with the Red Peak and that the history of Crow Mountain deposition actually begins with this unit. Moreover, in areas where the Alcova is missing because of nondeposition and the sub-Alcova sandstone has not been stained its characteristic color, the interval is not readily distinguished from the overlying Crow Mountain.

The transition from silty to sandy rocks in the upper part of the Red Peak probably represents a major change in the depositional history of Triassic strata in Wyoming. Picard (1967) believed the change in lithology was because of either uplift in the source area or the exposure of a new source area. Although both of these possibilities are tenable, other factors such as a lowering of base level or a subtle change in climate might have played a role in modifying conditions in the source areas. Moreover, studies by Blair and Bilodeau (1988) and Heller and others (1988) suggest that grain size can increase in depositional basins during periods of tectonic quiescence in the source area. Most geologists believe that the sub-Alcova sandstone was deposited in a nearshore marine environment (for example, Picard, 1978). The unit is probably a regressive deposit; however, it might be transgressive and thus represent the first deposits of the marine invasion that subsequently deposited the Alcova Limestone. The abrupt appearance of calcite cement in the sub-Alcova sandstone could support this conclusion. If the sub-Alcova is a transgressive unit, then it is more related to the Alcova than to either the Red Peak or the Crow Mountain and thus might better be included as a member of the Alcova. In contrast to a marine setting, Cavaroc and Flores (1991) postulated a nonmarine environment for this unit, citing the variegated color and turbation as evidence of pedogenic processes.

Late Early Triassic Transgression

Following deposition of the sub-Alcova sandstone, marine water from the Cordilleran miogeocline to the west moved rapidly eastward across a broad, almost flat surface as indicated by the great aerial extent and relative thinness of the Alcova Limestone. The marine invasion might have been caused by a general rise in sea level, a broad regional downwarp, or some combination of these processes. In any case, if a featureless marine shelf or coastal plain existed adjacent to the miogeocline, only a slight relative rise in sea level would have caused extensive transgression. Controversy exists as to whether or not any erosion preceded the transgression. The contact between the sub-Alcova sandstone and the Alcova Limestone is usually described as sharp and conformable; however, Picard (1967) thought the contact might be somewhat disconformable, and Hose (1955), in the central Bighorn Mountains, considered the contact erosional (this is in agreement with an observation I made in the southeastern Bighorn Mountains). Moreover, Hose, noting an apparent thinning of the Red Peak northward along the east flank of the Bighorn Mountains, stated that an angular unconformity might exist.

Marine Platform Carbonate Rocks

Conspicuously positioned near the middle of thousands of feet of Triassic clastic strata (fig. 8) is a thin carbonate unit, the Alcova Limestone, that extends over 31,000 mi^2 of eastern and central Wyoming (Picard, 1978). The unit was first described by Darton (1906) but remained unnamed until Lee (1927) applied the term "Alcova Limestone" to the rocks and assigned the unit as a member of the Chugwater Formation. The name Alcova comes from a small community on the northwestern flank of the Laramie Range where the rocks are particularly well exposed. When Love (1939) redefined the Chugwater Formation, the Alcova was not included because in the Absaroka Range, where Love subdivided the formation, the limestone is absent. Next, Branson and Branson (1941) raised the Chugwater to group status and assigned the unit, the Alcova Dolomite, as a formation of the group. When High and Picard (1967) redefined the overlying Crow Mountain Sandstone they included the Alcova a member of that formation (this reorganization of the Crow Mountain is not used by the U.S. Geological Survey).

The Alcova is usually a single, cliff-forming ledge of limestone about 15 ft thick (fig. 9). Recrystallization of carbonate mudstone has formed a hard, finely crystalline limestone. A calcium to magnesium ratio of greater than 50 and slightly negative O^{18} values led Cavaroc and Flores (1991) to postulate minimal diagenetic dissolution and cementation for limestones of the Alcova in the central Bighorn Mountains. The limestone is locally dolomitic, and Carini (1964), in his regional study of the Alcova, reported dolomite in the Alcova to be calcium-rich protodolomite and stated that outcrops of dolomitic Alcova are commonly marked by the presence of authigenic feldspar. In addition, along the eastern flank of the Bighorn Mountains, dolomitization appears to be somewhat structurally controlled, as noted by the increased amount of dolomitic rocks in areas that have been fractured by folding; this phenomenon apparently results from the introduction of magnesium-rich

Figure 8. Middle part of the Chugwater Group, southeastern Bighorn Mountains. 1, Paralic sandy rocks (Crow Mountain Sandstone). 2, Marine platform carbonate rocks (Alcova Limestone) (very thin bed). 3, Nearshore marine sandy rocks (sub-Alcova sandstone, Red Peak Formation). 4, Paralic silty rocks (upper platy facies, Red Peak Formation). 5, Paralic silty rocks (alternating facies, Red Peak Formation).

groundwater by way of these fractures. Marginal to the southwestern Powder River Basin, Storrs (1986) reported the Alcova to average 80 percent carbonate (chiefly calcite), the remaining 20 percent consisting mostly of silt-sized detritus. On the eastern flank of the Bighorn Mountains, McLellan (1968) reported that the clastic fraction tends to increase upward within the unit. Carini (1968) found the detrital fraction to contain quartz, feldspar, and mica. Heavy minerals include zircon, tourmaline, rutile, garnet, biotite, and chlorite; illite is the dominant clay mineral. Light gray is the most common basic color, but hematite staining has left most rocks mottled or banded pink, purple, or, more rarely, brown. Some rocks have been stained dark gray with oil, and these petroliferous rocks release a fetid order when struck. The rocks are usually very thinly laminated, either horizontal or wavy, and where wavy they have a crenulated appearance. Laminations commonly stand out in sharp relief because of differential weathering. Much of the lamination is apparently algal in origin. Bedding is thin to medium, and rare occurrences of small-scale cross-stratification, stylolites, and oscillation ripple marks have been reported. The unit weathers to slabby or platy fragments and is locally vertically jointed. Stromatolites are common, and most workers described them from the lower part of the unit. Cavaroc and Flores (1991), however, observed that stromatolites are restricted to the upper part of the unit in the Red Wall area in the south Bighorn Mountains.

Fossils are extremely rare in the Alcova. Specimens are limited to poorly preserved external casts and molds or calcite-filled voids of pelecypods and, even rarer, gastropods. The reader is referred to Darton (1906), Lee (1927), Branson and Branson (1941), Pipiringos (1953, 1957), and Carini (1964) for more information on the occurrence of mollusks in the Alcova. Fossil remains of a marine reptile were discovered near Casper, Wyoming, and described by Case (1936). At the time, *Corosaurus alcovensis* represented the only New World nothosaur described from

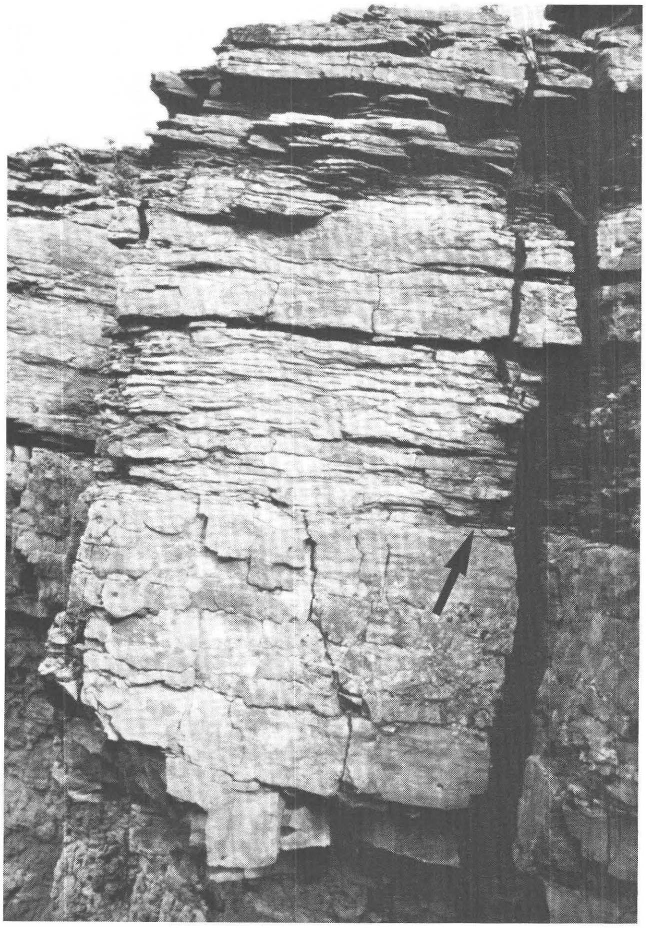


Figure 9. Ledge of the Alcova Limestone, southeastern Bighorn Mountains. Pencil (arrow) shown for scale.

North America. The contact of the Alcova with underlying rocks is sharp and probably conformable; the contact of the Alcova with the overlying Crow Mountain Sandstone is sharp and disconformable.

Where the Jelm Formation overlies the Alcova the contact can be sharp or gradational, but it is disconformable as well. Regionally, the Alcova is 3–15 ft thick but can be as thick as 30 ft (Picard, 1978). The thickest Alcova is reported in the general vicinity of Alcova, Wyoming. These differences in thickness, which are also shown on an isopach map by Storrs (1986), indicate that minor differential subsidence occurred on the platform. In some areas of eastern Wyoming, the limestone can locally thin to zero. At one outcrop in the central Bighorn Mountains I observed the unit to be represented by only a thin, discontinuous zone of limy, tabular sandstone clasts with rounded edges (fig. 10) as if representing the disturbed feather edge of the deposit. Rarely, the Alcova interval contains several separate limestone ledges. The Alcova is present in the northwestern Laramie Range and intermittently along the eastern flank of the Bighorn Mountains as far north as the Yellowtail Dam area in southern Montana. West of the Powder River Basin, the Alcova thickens with the addition of several limestone beds and extends almost to Jackson, Wyoming (Picard, 1967; High and Picard, 1969). Picard (1975) stated that the Alcova correlates with the sandstone and limestone unit at the top of the Thaynes Formation in westernmost Wyoming. The Alcova thins regionally toward the east, and a segment of its zero edge lies under the Powder River Basin. Although the unit can be recognized in the subsurface as far east as Upton, Wyoming, the Alcova is not present in the Black Hills. Some segments of the zero edge probably represent a paleo-shoreline and mark the maximum extent of the Alcova depositional basin. Absence by nondeposition is supported by some outcrop observations that reveal that where the limestone thins or pinches out the units above and below the formation remain unaffected. It is probable that to the east in the subsurface some of the regional thinning, and segments of the zero edge itself, resulted from beveling by Jurassic erosion. Thus, the original extent of the unit might have been more widespread than its present-day distribution indicates. On electric logs the Alcova is quite distinctive because of its prominent high resistivity response on both long and short normal curves; the spontaneous potential curve is usually unaffected.

The rare mollusks reported from the Alcova have all proven nondiagnostic in determining the age of the unit. And, according to Storrs (1986), the nothosaur genus *Corosaurus* reported from the Alcova had a worldwide distribution and a pan-Triassic age span; it is useless in determining the exact age of the unit. The Thaynes in western Wyoming contains Early Triassic fossils, and the Alcova is probably the same age; however, the sandstone and limestone unit at the top of the Thaynes, to which the Alcova has been directly correlated, contains no fossils useful for determining age. E.K. Maughan (written commun., 1992) stated that he has collected conodonts from the Alcova from the northern flank of the Uinta Mountains in northern Utah that were identified by B.R. Wardlaw (U.S. Geological Survey) as Early Triassic in age.

The Alcova was deposited on a broad, shallow marine platform that extended eastward from the Cordilleran miogeocline. Part of the eastern edge of this platform was in the area of the Powder River Basin, about 250 mi away from the main body of the sea. The presence of stromatolites in the Alcova indicates that the unit was deposited in warm, shallow, marine water under low-energy conditions. Evaporitic conditions are indicated by extreme positive O^{18} values and by the fact that the carbonate-precipitating water was hypersaline (Storrs, 1988). These evaporitic conditions could account for the apparent low diversity of fauna represented in the fossil record. Along with warm temperatures, hypersalinity requires an environment with little freshwater input from the landward

Figure 10. Discontinuous zone of limy sandstone clasts representing the disturbed depositional edge of the Alcova Limestone, east-central Bighorn Mountains. Pencil shown for scale.

side of the system; this is supported by the scarcity of clastic detritus in the limestone. Hypersalinity also requires limited access of water with normal-marine salinity entering the system from the seaward side. The absence of evaporites in the Alcova indicates, however, that some fresh or normal marine waters were able to enter the environment and maintain almost normal salinity. Most likely, the Alcova was deposited in a broad system of restricted bays and lagoons. On a more regional scale, Storrs (1986, 1988) suggested some type of physical barrier west of the Powder River Basin that would have limited normal marine circulation and resulted in the hypersaline conditions. Cavaroc and Flores (1991), in the southern Bighorn Mountains, used the upward progression from algal mats to algal heads to indicate deepening water. Because most workers have reported stromatolites to be restricted to the lower part of the unit and because crenulated laminations in the strata directly overlying the stromatolites are probably algal in origin, perhaps in these areas water depths shallowed with time. In either case, only one transgressive-regressive cycle is indicated. In the subsurface of the northern part of the Powder River Basin electric logs show the Alcova splitting into as many as six individual beds in a zone as thick as 150 ft. This phenomenon appears to represent conditions at the edge of the Alcova depositional basin and might indicate that the Alcova was deposited during several transgressive-regressive episodes. Berryman (1942) described 10–20 ft of gypsum at the top of the Chugwater beyond the established Alcova zero edge in an area south of Glenrock, Wyoming. This area is too far south for the gypsum to be associated with the Jurassic Gypsum Spring Formation, and the rocks might represent the most distal edge of the Alcova depositional environment. Otherwise, evaporites are not known to be present anywhere near the feather edge of the Alcova.

Post-Alcova Erosion

Most workers agree that a disconformity exists between the Alcova Limestone and the overlying Crow Mountain Sandstone. An unconformity at this level was first suggested by Thomas (1949). Pipiringos and O'Sullivan (1978) later designated it the Tr–2 and stated that it represented only a minor hiatus. What is debated is whether or not significant erosion occurred at this time. Limestone fragments, apparently from the underlying Alcova, in the basal few inches of the Crow Mountain have been reported by many workers (see Pipiringos and O'Sullivan, 1978), and their presence provides the best evidence that at least some erosion occurred at this level. Some actual scouring of the Alcova by the overlying Crow Mountain has been observed, but most workers have described it as minor. Tohill and Picard (1966) reported as much as 1 ft of relief in west-central Wyoming, but High and Picard (1967) reported that nowhere does the Crow Mountain completely truncate the limestone. In contrast to this, E.K. Maughan (written commun., 1992) noted that in an area of central Wyoming where the Alcova is missing a layer as thick as 2 ft containing limestone clasts is at the stratigraphic level of the Alcova, and he interpreted this as evidence of the erosional removal of the Alcova (see Maughan, 1972). Where I have observed a similar phenomenon in the central Bighorn Mountains, the debris could just as well be interpreted as the disrupted feather edge of the deposit. Cavaroc and Flores (1991) thought that the Crow Mountain might have cut completely through the Alcova in one area of the southern Bighorn Mountains, but these workers also observed that at some places in the area algal head mounds on the top surface of the Alcova indicate that significant subaerial erosion did not occur in these areas.

P.J. Lewis and H.D. Hadley (in Reeside and others, 1957) described a unit (discussed later in this report) at the top of the Spearfish in the Black Hills that has characteristics similar to the Crow Mountain, and they reported that the unit contains chert and carbonate clasts at its base. If the unit is a Crow Mountain equivalent, this layer might represent the Tr–2 in the Black Hills.

Late Early or Middle (?) Triassic Progradation

Following the westward withdrawal of marine water into the main body of the Cordilleran miogeocline and probable subsequent subaerial erosion in eastern and central Wyoming, clastic sediments once again prograded westward across the exposed cratonic platform. Unlike previous Triassic detritus that was dominated by silt, these sediments contained abundant sand. This implies increased uplift in the source area or exposure of an entirely new source area. Possibly, the withdrawal of marine water, subsequent erosion, and change in the source area were all related to a single tectonic event.

Paralic Sandy Rocks

Overlying the Alcova Limestone along the eastern flank of the Bighorn Mountains is a sequence of cliff-forming, salmon-colored sandy rocks. Similar rocks make up most of a sequence of strata that is present along the southern margin of the Absaroka Range between the Red Peak Formation (the Alcova is not present in this area) and the overlying Upper Triassic Popo Agie Formation. Love (1939) named this interval of rocks the Crow Mountain Sandstone Member of the Chugwater Formation. He included strata at the top that most later workers considered a separate unit (the unnamed red beds). The name Crow Mountain was raised to formational status when Branson and Branson (1941) designated the Chugwater a group. Tohill and Picard (1966) divided the Crow Mountain into two units, a basal sandstone unit and an upper sandstone and siltstone unit. Next, High and Picard (1967) included two lower units to the Crow Mountain, the Alcova Limestone Member (the Alcova Limestone, a formation rank unit this report) and the underlying variegated sandy facies (the sub-Alcova sandstone of this report). None of the changes proposed by High and Picard (1967) are used by the U.S. Geological Survey, but the basal sandstone and upper sandstone and siltstone units of Tohill and Picard (1966) are quite useful in describing the Crow Mountain in the Bighorn Mountains (see also Maughan, 1972).

The basal sandstone part of the Crow Mountain is composed of well-sorted, calcareous sandstone. The sand is mostly very fine grained, but fine, medium, and even rare coarse grains are present. Locally, the unit coarsens upward. Grains are subrounded to well rounded, and some larger grains appear frosted. Tohill and Picard (1966) reported the sandstone to be a subarkose. The unit commonly weathers a distinctive orangish red (locally light gray) and is locally friable. The unit is massive to thick bedded, but thinner beds are present in the uppermost part. Medium-scale, trough cross-stratification is locally well developed; tabular cross-stratification is less common. Straight-crested, asymmetrical ripples marks are present on bedding surfaces in the uppermost part of the unit. Burrows and bioturbation are present, but fossils are absent. This part of the Crow Mountain typically crops out as a smooth, almost vertical cliff, and along the southeastern flank of the Bighorn Mountains it makes up the higher part of the Red Wall above the Alcova Limestone (fig. 8). Regionally, the basal sandstone thins toward the southwest (Tohill and Picard, 1966), and thicknesses vary between 16 and 130 ft (Picard, 1978). Along the eastern flank of the Bighorn Mountains, McLellan (1968) reported that the unit is 37–94 ft thick. Sandstone in the basal part of the Crow Mountain is porous and permeable, and oil staining is common. The formation is known to produce oil in at least five oil fields in eastern and central Wyoming. Picard (1978) reported that oil from the unit is of low gravity and is high in sulfur.

The upper sandstone and siltstone part of the Crow Mountain is thinner and somewhat less resistant than the basal sandstone. Sandstone is very fine grained, well sorted, calcareous, and orangish red or light gray; individual grains are well rounded. Sedimentary structures include parallel bedding, small-scale trough cross-stratification, and straight-crested, asymmetrical ripple marks. Siltstone is well sorted, calcareous, and red or yellowish gray; individual grains are subangular to subrounded, and the rock is classified a subarkose (Tohill and Picard, 1966). Claystone partings are present locally, and Tohill and Picard (1966) reported claystone-pellet conglomerate and rare mud cracks. The upper sandstone and siltstone part of the Crow Mountain is laterally variable because of facies changes. As with the lower part of the Crow Mountain fossils are absent. The contact between the two subdivisions of the Crow Mountain is gradational and conformable, and a slight change in color and grain size helps to distinguish the two subdivisions. The contact of the upper sandstone and siltstone unit with overlying rocks is sharp and disconformable. Tohill and Picard (1966) reported the thickness of the unit to range from 1.5 to 19 ft in the southern Bighorn Mountains, and thicknesses reported by McLellan (1968) for the eastern flank of the mountains are within this range.

The Crow Mountain probably represents a depositional environment in which energy decreased with time, probably because of shoaling. Initial deposition was probably in a shallow-marine setting. The basal sandstone part of the formation likely represents a regressive shoreface sequence as indicated by the upward increase in grain size and by the upward decrease in the scale of sedimentary structures. Because of the presence of frosted sand grains in this part of the Crow Mountain, an eolian component to the depositional history cannot be ruled out. Decreasing water depth culminated with deposition of the upper sandstone and siltstone unit. Tohill and Picard (1966) presented evidence that suggests a beach or tidal-flat environment for this part of the formation. Tohill and Picard also postulated that the source areas for Crow Mountain sediment were to the east and north because, regionally, the formation thickens and becomes coarser grained in those directions. These workers went on to say that this configuration, coupled with paleocurrent measurements, suggests a northwest-trending coast line.

The Crow Mountain is not listed among those units present in the Black Hills, and most geologists believe that the formation has been removed by Jurassic erosion in this area. However, J.D. Love (in Reeside and others, 1957) reported that rocks similar to the Crow Mountain are in the upper part of the Spearfish Formation. This statement coincides with descriptions by Imlay (1947) of rocks in the upper part of the Spearfish at two localities in the southern Black Hills and with the description of unit 4 of the Spearfish as defined by P.J. Lewis and H.D. Hadley (in Reeside and others, 1957). It is thus possible that where the Alcova Limestone is missing in the Black Hills because of nondeposition rocks equivalent to the Crow Mountain of central Wyoming exist unrecognized at the top of the Spearfish. West of the Bighorn Mountains, the Crow Mountain extends into western Wyoming, and in the westernmost part of the state it grades into the lower part of the Ankareh Formation (fig. 7).

Rocks in south-central Wyoming stratigraphically equivalent to the Crow Mountain are present in the lower part of the Jelm Formation, and they represent a depositional environment distinctively different from that of the Crow Mountain. The Jelm was named by Knight (1917) for rocks exposed near the eastern base of Jelm Mountain in the Laramie Basin of southern Wyoming. Prior to this, these rocks had been included in the upper 250 ft of the Chugwater Formation (the Alcova Limestone is not present in this area, so the Jelm rests on the Red Peak Formation of this report). Pipiringos (1957) later designated Red Mountain, located near the Colorado-Wyoming State line about 5 mi southeast of Jelm Mountain, as a better type section. Based on good exposures in south-central Wyoming, Pipiringos (1968) next defined the Jelm as those rocks between the Alcova Limestone and the Popo Agie Formation and designated the lower part of the Jelm the Red Draw Member and the upper part the Sips Creek Member. The Red Draw at its type section is about 100 ft thick and consists mostly of reddish-brown shale, siltstone, and sandstone. The Sips Creek at its type section is about 315 ft thick and consists of white or orange, cross-stratified sandstone in the lower part and reddish-brown siltstone in the upper part. Fossils are not reported from either member. A major disconformity separates these two members, and Pipiringos and O'Sullivan (1978) correlated this surface with the unconformity at the top of the Crow Mountain of Tohill and Picard (1966). Thus, the Crow Mountain of central Wyoming is a distant lateral facies of the Red Draw Member of the Jelm in south-central Wyoming.

While the Crow Mountain was being deposited in central Wyoming following the post-Alcova erosional event, sediments now contained in the Red Draw were accumulating to the south. The source of Red Draw sediments was probably different than that of Crow Mountain sediments; perhaps it was older Triassic rocks exposed in the ancestral Rocky Mountains to the south. The unit is probably nonmarine. Sediments possibly accumulated on a coastal plain that might have contained fluvial, lacustrine, and minor eolian subenvironments (see Picard, 1978; Peterson, 1988); however, shallow-marine conditions might have existed to the west. The northward transition from Jelm lithology to Crow Mountain lithology is difficult to study because of the lack of exposures in critical areas. In addition, lithofacies of both units vary in their regional extent, causing a broad interfingering relation, such that simply drawing a line separating the two formations is troublesome. Because of this, a problem exists in naming rocks above the Alcova and below the Jurassic Sundance Formation adjacent to the southern margin of the Powder River Basin in the northern Laramie Range. Published studies of these rocks, primarily theses, contain descriptions that could match parts of either the Jelm or the Crow Mountain.

Post-Crow Mountain Erosion

In central Wyoming a disconformity separates the Crow Mountain Sandstone from overlying rocks (Tohill and Picard, 1966). The disconformity exists within the Crow Mountain of Love (1939) because when Love defined the Crow Mountain he included strata at the top that most workers now consider a separate unit (the unnamed red beds). Relief on this surface is less than 2 ft in the southern Bighorn Mountains (McLellan, 1968), but in other areas to the west it can be as much as 10 ft (Picard, 1978). In south-central Wyoming a disconformity separates the Red Draw and Sips Creek Members of the Jelm Formation (Pipiringos, 1968). Pipiringos and O'Sullivan (1978) have determined that these two surfaces are the same and have designated it the Tr–3 unconformity. The Tr–3 represents a major event in Triassic history, and can be traced throughout the Western Interior. Pipiringos and O'Sullivan (1978) estimated that the erosional event associated with the Tr–3 lasted a million years.

The Missing Middle Triassic

Very few fossils of Middle Triassic age have been identified in the Western Interior; the closest occurrence to the Powder River Basin is in east-central New Mexico. Middle Triassic fossils are relatively common in west-central Nevada. Whether or not unfossiliferous rocks of Middle Triassic age are present in the middle and northern Western Interior is the subject of debate. In the Colorado Plateau, the Upper Triassic Chinle Formation is separated from the overlying Lower Triassic Moenkopi Formation by the Tr–3 unconformity (Pipiringos and O'Sullivan, 1978). Some workers believe that the uppermost part of the Moenkopi might be Middle Triassic (Stewart and others, 1972; Lucas and others, 1985, Blakey, 1989). If Middle Triassic strata are present in central Wyoming, they would be below the Tr–3 unconformity because rocks above this surface are all considered to be Upper Triassic. Candidates include some or all of the Crow Mountain or the Red Draw Member of the Jelm Formation. The next question is: were sediments of Middle Triassic age simply never deposited, or were they removed (or partly removed) by erosion? And, if they were removed by erosion, was it by the Tr–2 event or the Tr–3?

Late Triassic Deposition

Following the Tr–3 erosional event, deposition resumed throughout the Western Interior. The source areas and depositional settings were apparently quite different, however, as compared to those of rocks deposited before the erosional event.

Nonmarine Clastic Rocks

Resting unconformably on the Tr–3 erosion surface in central Wyoming is a sequence of rocks known as the "unnamed red beds" (High and Picard, 1965). West of the Powder River Basin the unit is 35–100 ft thick, and it is composed mostly of reddish siltstone with some massive to cross-stratified, very fine grained sandstone and mudstone. Along the eastern flank of the Bighorn Mountains the unit consists of siltstone interbedded with lesser amounts of mudstone and clay shale. Isolated sandstone bodies are common in the lower part of the unit, and, locally, they are the dominant lithofacies. Most rocks are calcareous. The siltstone is usually a shade of red. The rocks are structureless, to parallel or ripple laminated; some appear bioturbated. Small-scale, trough cross-stratification is common. Bedding thickness varies from very thin to thick, and plant fragments are present rarely on bedding planes. Some erosional surfaces are present within siltstone sequences. The mudstone is usually a shade of red, and the clay shale is usually greenish gray. The sandstone is in lenticular packages, as thick as 40 ft, that commonly are composed of amalgamated bodies separated by surfaces of erosion (fig. 11). These rocks are usually some shade of red, very fine to fine grained, and well to poorly sorted. Dark, argillaceous laminae commonly define horizontal bedding or medium- to small-scale, trough cross-stratification. The lenticular packages have sharp scour bases with overlying argillaceous rip-up clasts; the upper contact can be sharp or gradational. Ripple laminations and ripple marks are common especially near the top. Tabular bodies, usually less than 10 ft thick, composed of silty, very fine grained sandstone also are present isolated in the siltstone. These rocks are ripple laminated or bioturbated. In higher parts of the unnamed red beds, sandy to silty, conglomeratic limestone beds are present locally. And, Cavaroc and Flores (1991) noted the occurrence of one bed of gypsum in the upper part of the unnamed red beds in the southern Bighorn Mountains. I know of no identifiable fossils reported from the unnamed red beds; however, based on stratigraphic position, the unit is almost certainly Late Triassic in age and is part of the

Figure 11. Lenticular sandstone package in the unnamed red beds unit showing individual channel-form bodies and well-developed trough cross-stratification, southeastern Bighorn Mountains.

Chugwater Group. It unconformably overlies the Crow Mountain Sandstone and, in the area of the Powder River Basin, is unconformably overlain by Jurassic strata. Because the top of the unnamed red beds unit has been beveled by Jurassic erosion, the original thickness of the unit in the area of the Powder River Basin will never be known; however, it is at least 240 ft thick in the southern Bighorn Mountains.

In the past, the unnamed red beds unit in central Wyoming has been included in the upper 36 ft of the Crow Mountain (Love, 1939); isolated as a separate unit in the Chugwater Group between the Crow Mountain and the Popo Agie (High and Picard, 1965); included as the lowest part of the Popo Agie Formation (High and Picard, 1967); or referred to as the Jelm Formation (Picard, 1978). Pipiringos and O'Sullivan (1978) correlated the unnamed red beds of central Wyoming with the Sips Creek Member of the Jelm Formation of south-central Wyoming (fig. 7). If the Sips Creek has a wider areal distribution than the underlying Red Draw Member of the Jelm, then in areas where the Red Draw has, by lateral facies change, become the Crow Mountain, part of the Jelm can overly the Crow Mountain. Based on this, substituting the name Jelm for the term "unnamed red beds" on the eastern flank of the Bighorn Mountains might be justified. But, using the name Jelm implies all of the Jelm, which is not the case. Moreover, it is possible that in the southeastern Bighorn Mountains the main body of the unnamed red beds represents the upper part of the Jelm but that higher rocks in the unnamed red beds unit could be a part of the Popo Agie Formation. Williston (1904) named the Popo Agie for exposures on the northeastern side of the Wind River Range. High and Picard (1965) divided the Popo Agie of central Wyoming in ascending order as follows: lower carbonate unit, purple unit, ocher unit, and upper carbonate unit. (High and Picard (1967) later added the unnamed red beds to the base of the formation.) The lower carbonate unit is 5–31 ft thick and consists of limestone microconglomerate, sandy and silty limestone, and silty dolomite; the contact with the underlying unnamed red beds unit is gradational through several feet. It is tempting to think of the sandy to silty, conglomeratic limestone reported in the upper part of the unnamed red beds unit in the southern Bighorn Mountains as an eastward extension of the lower carbonate unit of the Popo Agie; however, High and Picard (1965), in their original description of the unnamed red beds unit in areas where the unit is overlain by the lower carbonate unit of the Popo Agie, did mention that limestone microconglomerate also is present in the unnamed red beds unit in at least one area. So, the limestone observed in the unnamed red beds unit in the southern Bighorn Mountains is not totally unexpected. Moreover, most workers have noted that the eastern edge of the Popo Agie strata was beveled by Jurassic erosion and that the formation is not represented in the Bighorn Mountains. Whether or not the eastern depositional zero edge was this far east will probably never be known.

The best evidence that the unnamed red beds are nonmarine comes from the nature of the sand bodies in the lower part of the unit. Lenticular sandstone bodies with erosional bases and sedimentary structures that indicate an upward decrease in depositional energy almost certainly represent fluvial channel deposits. Tabular sandstone bodies represent crevasse-splay or sheet-flood deposits. Interchannel areas are now represented by the siltstone, mudstone, and shale that form the matrix of the unnamed red beds. The isolated limestone beds in the higher parts of the unit were probably deposited in shallow, freshwater lakes. These environments existed on a broad, flat, surface that might have been part of a westward-prograding coastal plain.

Top of the Triassic

Throughout the Western Interior, Triassic and Jurassic strata are separated by an erosional unconformity. In the Black Hills, northern Laramie Range, and southern Bighorn Mountains, Triassic rocks are separated from the Jurassic Sundance Formation by the J–2 unconformity (Pipiringos and O'Sullivan, 1978); in the central Bighorn Mountains, Triassic rocks are separated from the Jurassic Gypsum Spring Formation by the J–1 unconformity. North of the Wyoming-Montana State line, in the northern Bighorn Mountains, Triassic rocks are separated from the Jurassic Piper Formation by the J–2 unconformity, and, farther north, in the subsurface of the Williston Basin, Triassic rocks are thought to be separated from the Jurassic Nesson Formation by the J–1 unconformity. In the Bighorn Mountains the highest Triassic unit, the unnamed red beds, has been beveled by the Jurassic erosion, and the amount of missing rock is unknown. West of the Powder River Basin, rocks equivalent to the unnamed red beds are overlain by as much as 130 ft of the Popo Agie Formation (High and Picard, 1965). This unit thins toward the east because of Jurassic erosional beveling, and it has probably been completely removed in the Bighorn Mountains. The natural thickness of the Popo Agie probably decreases toward the east by depositional thinning, and thus the unit might not have been deposited in the area of the Powder River Basin.

TECTONIC INFLUENCE ON TRIASSIC SEDIMENTATION

As pointed out by S.S. Oriel (in McKee and others, 1959), isopach maps of total Triassic strata in Wyoming show a correlation between thick Triassic sections and anticlinal axes and between thin Triassic sections and post-Triassic structural basins. Greer (1985) reported that the same phenomena is apparent on isopach maps of the Goose Egg Formation. Moreover, Greer mentioned that a regional rectangular pattern could be detected on these maps. Triassic depocenters were probably defined by differential movement along regional zones of weakness in the Precambrian basement. Following the Triassic, renewed movement along these zones in a reverse sense resulted in the observed paradox.

SUMMARY

1. During the Late Permian and the earliest Triassic, the area that is now the Powder River Basin was on the east side of a large marine basin. Clastic sediments sourced from the east and possibly the south were deposited in nonmarine environments on an almost flat surface under low-energy conditions. Episodically, rapid marine transgressions deposited eastward-thinning tongues of carbonate and evaporite rocks. The Triassic part of this history is now represented by the siltstone, limestone, and gypsum deposits included in the upper part of the Goose Egg Formation and in equivalent strata in the lower part of the Spearfish Formation.

2. During the Early Triassic, a large volume of clastic sediments similar to those deposited earlier in the period prograded westward and overwhelmed the depositional setting. Exactly how these sediments were deposited is debated, but they probably accumulated under fluvial, tidal, and shallow-marine conditions. This great bulk of red beds is now contained in the Red Peak Formation and the middle and probably upper parts of the Spearfish.

3 Later in the Early Triassic, following the accumulation of red beds, a physical change in the source area (or the introduction of a new source area), coupled with an increase in depositional energy, resulted in an accumulation of sandy sediments. These paralic sandstone beds are now included in the top of the Red Peak.

4. In the late Early Triassic a rapid marine transgression spread marine water over a vast area of central Wyoming. The Alcova Limestone now contains the shallow-water deposits of this event.

5. Deposition of the Alcova ceased as marine water regressed from the area. What imminently followed this event is uncertain, but a brief period of nondeposition occurred that resulted in the Tr–2 unconformity.

6. In the latest Early Triassic, or possibly earliest Middle Triassic, sandy sediments prograded westward across the Alcova Limestone. These shoreface deposits now make up the Crow Mountain Sandstone.

7. Much of the record of Middle Triassic time was lost during a period of erosion or nondeposition that followed Crow Mountain deposition. The Tr–3 unconformity is of regional extent, and it represents the most significant erosional event of the Triassic Period in the Western Interior.

8. During the Late Triassic, clastic sediments prograded westward across the Tr–3 surface depositing the unnamed red beds unit under nonmarine conditions.

9. Deposition might have continued in the area of the Powder River Basin until the close of the period, but any uppermost Triassic rocks that were deposited have been lost to Jurassic erosion.

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