SANDSTONE DETRITAL MODES, CENTRAL UTAH FORELAND REGION: STRATIGRAPHIC RECORD OF CRETACEOUS-PALEOGENE TECTONIC EVOLUTION[1]

DAVID BLOOM

WILLIAM R. DICKINSON
Laboratory of Geotectonics
Department of Geosciences
University of Arizona
Tucson, Arizona 85721

TIMOTHY F. LAWTON AND KERRY F. INMAN
SOHIO Petroleum Company
5400 LBJ Freeway
Dallas, Texas 75240

ABSTRACT: Within a sedimentary basin, the analysis of sandstone petrofacies defined by detrital modes of framework grains is a powerful means to detect areal or temporal changes in provenance related to evolving paleotectonics and paleogeography. In the central Utah foreland, the regional tectonic transition from thin-skinned Sevier thrusting during the Late Cretaceous to deep-seated Laramide deformation during the Paleogene is recorded by four contrasting petrofacies in uppermost Cretaceous and Paleogene sandstones of fluvial and deltaic facies. Coeval subaqueous environments were marine during the Cretaceous and lacustrine during the Paleogene. The Cretaceous quartzose petrofacies (~Qm93, F1, Lt6) of mainly Campanian age was derived from sedimentary sources in the active Sevier thrust belt, and transported into a broad adjacent foreland basin by transverse paleocurrents. The heterogeneous Cretaceous quartzolithic petrofacies (~Qm40-70, F5-20, Lt25-40) of uppermost Campanian age was derived from the Sevier orogenic belt during the waning phases of Sevier tectonism, and was distributed by longitudinal paleocurrents within a foreland basin whose floor had begun to be deformed by incipient Laramide tectonism. The Paleogene feldspathic petrofacies (~Qm50, F35, Lt15) in Paleocene and Eocene strata is composed of arkosic detritus that was derived principally from basement rocks then exposed in the extensive San Luis uplift of Laramide age in south-central Colorado. The feldspathic sand was transported to the Laramide Uinta Basin and vicinity in Utah by a trunk river which flowed generally northwestward between monoclinal Laramide uplifts of the Colorado Plateau. The Paleogene quartzolithic petrofacies (~Qm75, F2, Lt23) of dominantly Eocene age was derived mainly from sedimentary sources in the Laramide Uinta uplift to the north, but also in part from similar sources in the dormant Sevier thrust belt to the west. In the Rocky Mountain and intermountain regions, more widespread and systematic use of petrofacies analysis, in combination with standard paleocurrent and lithofacies analyses, could lead to better definition of the timing of Laramide uplift and improved understanding of the evolution of complex Laramide paleogeography.

INTRODUCTION

Within the Mesozoic-Cenozoic forearc region along the Pacific margin of the United States (Fig. 1), compilations of sandstone detrital modes determined by the point counting of thin sections are used extensively to define stratigraphic petrofacies (Dickinson and Rich 1972) controlled by provenance relations. The framework constituents of literally thousands of sandstones have by now been reported in the literature in the form of either tabulated individual counts, triangular compositional plots, or calculated suite means. Knowledge of petrofacies patterns is used to establish paleogeographic settings and to infer paleotectonic environments (Ingersoll 1983). Recent topical studies show how this petrofacies methodology can effectively separate detrital contributions from mixed sources (Johnson 1984), detect successive stages in basin evolution (Pacht 1984), and date significant tectonic transitions (Heller and Ryberg 1983).

Within the Cordilleran foreland region farther inland (Fig. 1), petrofacies methodology has thus far not been used as systematically to constrain inferences about evolving paleogeography and changing paleotectonics. This remains the case despite encouraging results from promising regional studies in southwest Wyoming (Surdam and Stanley 1980) and southwest Montana (Suttner et al. 1981; Schwartz 1982). In this paper, we present an analysis of detrital modes for Upper Cretaceous and Paleogene sandstone suites from the central Utah foreland (see Fig. 1) to illustrate how petrofacies methodology can help answer key questions about foreland tectonics. We rely principally upon our own petrographic reconnaissance of the region, but also compare our data to results obtained by previous workers. A summary of most existing petrographic information was recently assembled by Bruhn et al. (1983, table 2).

Previous tectonic syntheses indicate that an extensive Cretaceous foreland basin occupied most of the present Colorado Plateau and much of the Rocky Mountain region east of a backarc thrust system (Fig. 1), which included the Sevier orogenic belt of Nevada and Utah, the Idaho-Wyoming thrust belt, and the Montana disturbed belt. This broad foreland basin was strikingly asymmetric, with an elongate western depocenter (or "foredeep") adjacent to the tectonic load of the thrust sheets (Jordan 1981; Lawton and Mayer 1982). Beginning in latest Cretaceous time and culminating during the Paleogene, Laramide deformation disrupted the previously integrated foreland basin to produce a geographically complex province of basement-cored and thrust-bounded uplifts (Fig. 1), which were separated by numerous local basins. In this paper, we focus on the effects of the tectonic transition from Sevier thrusting to Laramide deformation on sandstone detrital modes in the central Utah foreland.

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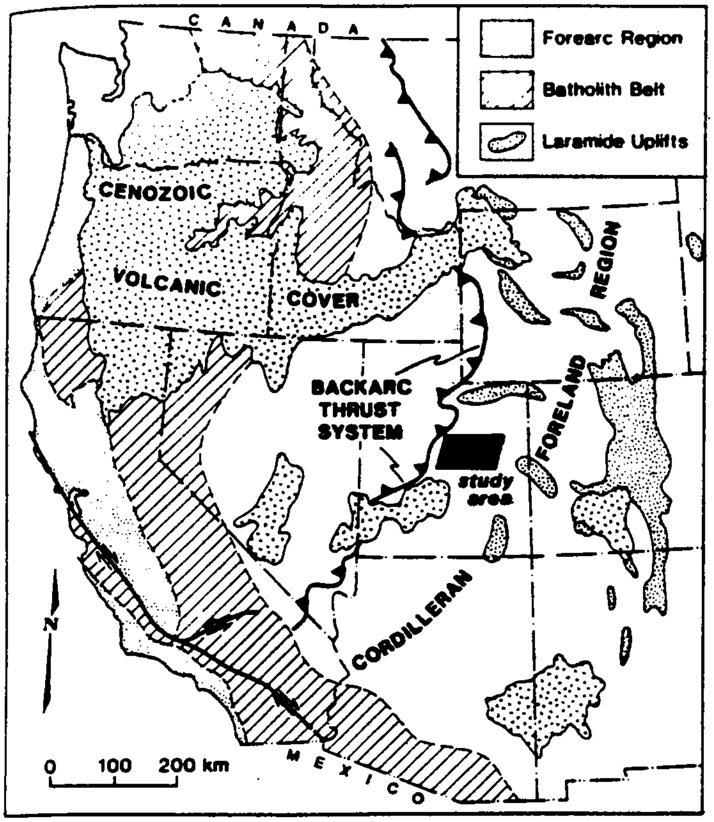

FIG. 1.—Geotectonic sketch map of western Cordillera of United States showing location of study area within foreland region east of backarc thrust system; batholith belt is locus of major Mesozoic plutons; Laramide uplifts farther inland formed largely in Paleogene time; forearc region along coast includes various forearc basins, subduction complexes, and suspect terranes developed along or accreted against the continental margin since mid-Late Jurassic time.

MODAL CONVENTIONS

The following classification of framework grain types is used for discussions of detrital modes:

a) Total quartzose grains (Qt) are the sum of monocrystalline quartz grains (Qm) and polycrystalline quartzose (or chalcedonic) lithic fragments (Qp), commonly chert.

b) Feldspar grains (F) include both plagioclases (P) and potassium feldspars (K) in varying proportions.

c) Unstable (siliciclastic) lithic fragments (L) are the sum of volcanic-metavolcanic (Lv) and sedimentary-metasedimentary (Ls) varieties, exclusive of quartzose-chalcedonic (Qp) types (chert, metaquartzite, etc.) and detrital limeclasts (Lc).

d) Total siliciclastic lithic fragments (Lt) are the sum of unstable (L) and quartzose (Qp) varieties.

For tables and diagrams, the proportions of quartzose grains, feldspar grains, and siliciclastic lithic fragments are recalculated to 100 percent, exclusive of other constituents. Summary detrital modes are then reported as QtFL and QmFLt populations (e.g., Qt50-F25-L25 or Qm40-F25-Lt35, etc.). In our petrographic work, quartz and feldspar crystals within lithic fragments, but individually larger than the lower limit of the sand range (i.e., 0.0625 mm), have been summed as quartz (Qm) and feldspar (F) grains (e.g., Ingersoll et al. 1984), but this convention may not have been followed by other workers whose data we cite.

Modal summaries thus follow standard practice, except that detrital limeclasts (Lc) are not included here with siliciclastic sedimentary and metasedimentary lithic fragments (Ls). In effect, calclithite is thus regarded as a variety of calcarenite, rather than litharenite. Most other workers have taken the opposite approach, by treating extrabasinal limeclasts (Lc) as simply a variety of the general category of sedimentary lithic fragments (Ls); in some studies, this alternative procedure may clarify provenance relationships (e.g., Mack 1984). However, we have found that the content of detrital limeclasts (Lc) within sandstone suites of central Utah is so variable, and seemingly capricious (see below), as to obscure fundamental compositional relationships if detrital limeclasts (Lc) are summed with the dominant siliciclastic grain types. Consequently, we prefer to report the content of detrital limeclasts separately, as percent of total framework grains, much as the content of mica (M) is customarily reported in detrital modes. We are uncertain whether the unsystematic distribution of detrital limeclasts (Lc) in the rocks we have studied reflects variable weathering of carbonate source rocks or variable behavior of carbonate framework grains during diagenesis. We observe, however, that relative proportions of siliciclastic grain types conform to patterns that are stratigraphically systematic, whereas variations in the content of detrital limeclasts appear to be essentially random.

STUDY AREA

Laramide deformation within the foreland region of the western United States caused structural disruption of the regionally integrated Cretaceous foreland basin. Pre-Tertiary strata were eroded from the crests of prominent Laramide uplifts, and Cretaceous beds were locally buried beneath thick Paleogene deposits in isolated Laramide basins. Eastward from the Sevier thrust front, the lateral outcrop continuity of synorogenic strata spanning the Cretaceous-Tertiary boundary is preserved only in central Utah along the eroded southern flank of the Uinta Basin (Fig. 2). To study the petrofacies record of the transition from Sevier to Laramide tectonics, our attention was thus drawn to the central Utah foreland. Within the study area (Fig. 3), Cretaceous and Paleogene strata are magnificently exposed from the Wasatch Plateau eastward along the line of the Book Cliffs (Cretaceous) and the Roan Cliffs (Paleogene) to Gray Canyon (Cretaceous) and Desolation Canyon (Paleogene) of the Green River.

STRATIGRAPHIC SETTING

Figure 4 depicts general stratigraphic and facies relations of outcropping strata in east-west transect across the study area. The diagram essentially represents an update of the pioneering work of Spieker (1946). As the

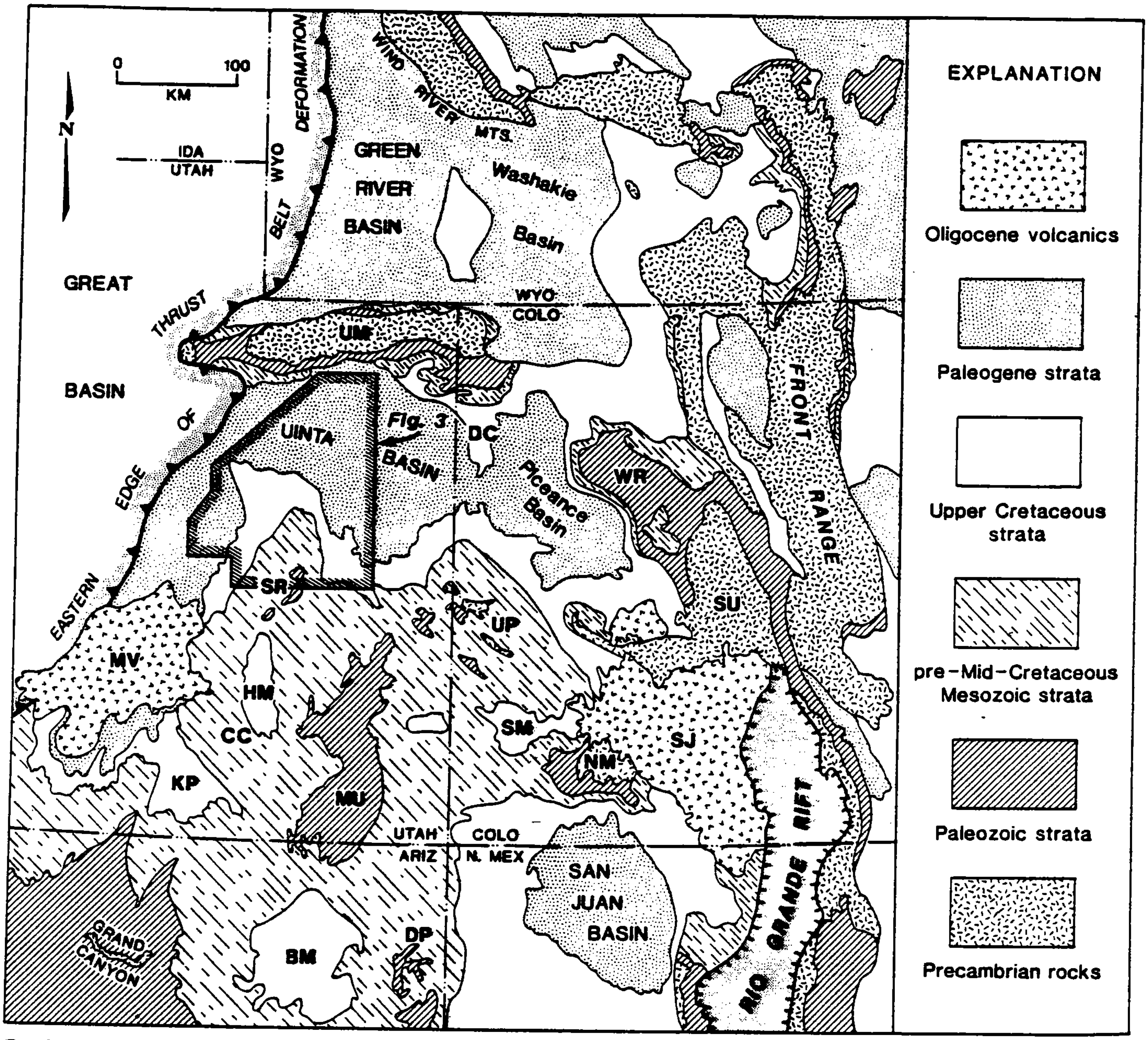

FIG. 2.—Geologic sketch map showing location of study area (Fig. 3) in central Utah foreland in relation to Laramide uplifts and basins of Colorado Plateau and Rocky Mountains. Key morphologic features: BM, Black Mesa; CC, Circle Cliffs uplift; DC, Douglas Creek Arch; DP, Defiance Plateau; HM, Henry Mountains region; KP, Kaiparowits Plateau; MU, Monument upwarp; MV, Marysvale volcanic field; NM, Needle Mountains uplift; SJ, San Juan volcanic field; SM, San Miguel Mountains; SR, San Rafael Swell; SU, Sawatch uplift; UM, Uinta Mountains; UP, Uncompahgre Plateau; WR, White River uplift.

strata dip gently northward into the Uinta Basin, younger horizons shown actually lie as much as 100 km north of the oldest horizons shown (see Fig. 3). Consequently, significant facies changes that occur in the subsurface of the Uinta Basin (e.g., Fouch 1975, 1976; Fouch et al. 1976; Ryder et al. 1976) are not adequately displayed by the diagram.

Cretaceous Units

Strata of the Cretaceous foreland basin are dominantly coarse-grained nonmarine rocks near the thrust front and finer-grained marine rocks east of the Wasatch Plateau (Fig. 3). The Cretaceous Indianola Group accumulated to a maximum thickness of about 4,000 m along the

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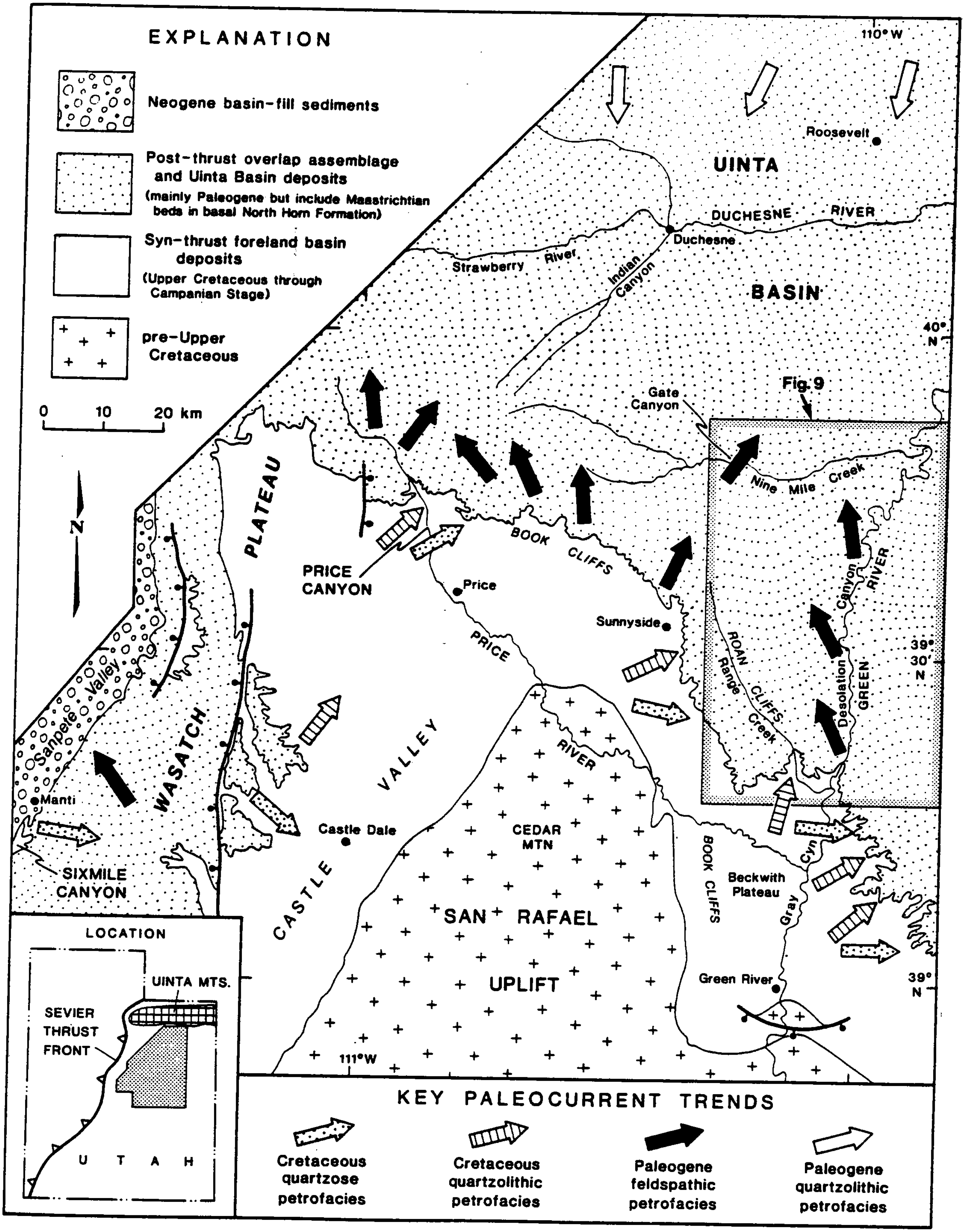
FIG. 3.—Geologic sketch map of central Utah foreland (adapted from Hintze 1980); paleocurrent data after Andersen and Picard (1972, 1974), Chapman (1982), Zawiskie et al. (1982), Lawton (1982, 1983a, 1986), and Figure 9; see Figure 2 for location. Balls on downthrown sides of normal faults (heavy lines).


EXPLANATION
Neogene basin-fill sediments
Post-thrust overlap assemblage and Uinta Basin deposits
(mainly Paleogene but include Maastrichtian beds in basal North Horn Formation)
Syn-thrust foreland basin deposits
(Upper Cretaceous through Campanian Stage)
pre-Upper Cretaceous
0 10 20 km
N
110° W
Roosevelt
UINTA
DUCHESNE RIVER
Duchesne
Strawberry River
Indian Canyon
BASIN
40° N
Gate Canyon
Fig. 9
Nine Mile Creek
PLATEAU
BOOK CLIFFS
PRICE CANYON
Price
Sunnyside
PRICE
ROAN CLIFFS
Range Creek
Canyon
Desolation
GREEN RIVER
39° 30' N
Sanpete Valley
WASATCH
CASTLE VALLEY
RIVER
Manti
Castle Dale
CEDAR MTN
BOOK CLIFFS
Beckwith Plateau
Gray Cyn
SIXMILE CANYON
SAN RAFAEL
UPLIFT
Green River
39° N
111° W
LOCATION
UINTA MTS.
SEVIER THRUST FRONT
U T A H
KEY PALEOCURRENT TRENDS
Cretaceous quartzose petrofacies
Cretaceous quartzolithic petrofacies
Paleogene feldspathic petrofacies
Paleogene quartzolithic petrofacies

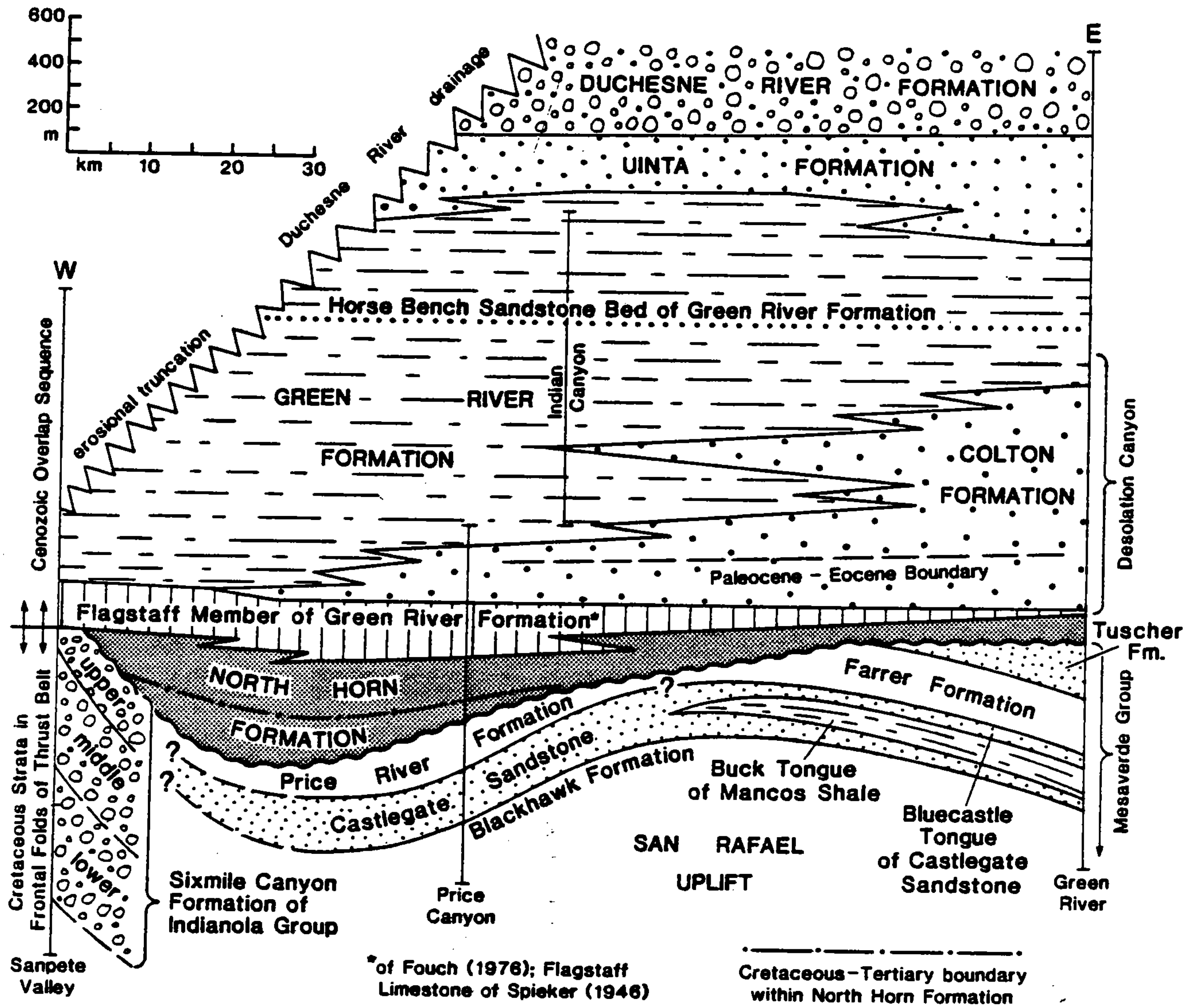


FIG. 4.—Schematic fence diagram of exposed stratigraphic units (subsurface relations not shown) across central Utah foreland showing general stratigraphic relationships of uppermost Cretaceous (mid-Campanian to Maastrichtian) and overlying Paleocene-Eocene strata in east-west profile (see Fig. 3 for locations of Sanpete Valley, Price Canyon, Duchesne River, and Green River). Adapted after Andersen and Picard (1972, 1974), Fouch et al. (1976, 1982, 1983), Ryder et al. (1976), Zawiskie et al. (1982), and Lawton (1982; 1983a, b; 1985, 1986).

deformed flank of the basin in central Utah (Spieker 1946; Jefferson 1982). Indianola sedimentation, which began in mid-Cretaceous time and continued through most of the Late Cretaceous (Lawton 1985), was coeval with the emplacement of Sevier thrust sheets, whose eastward migration eventually involved the synorogenic Indianola strata in thrust-related deformation. Near Sanpete Valley at the western edge of the study area (Fig. 3), Indianola beds are involved in the frontal anticline of a complex "triangle zone" that forms the leading edge of the Sevier thrust belt (Standlee 1982; Lawton 1983b, 1985). Conglomeratic Indianola strata dip homoclinally eastward (Fig. 4) off the frontal fold structure into the subsurface under the Wasatch Plateau. Indianola lithofacies patterns define a coarsening-upward progradational sequence dominated by nonmarine deposits of alluvial fans, fluvial braidplains, and sandy meanderbelts. These terrestrial lithofacies intertongue eastward and downward with deltaic and lagoonal successions, which grade in turn to sequences of shoreface and delta-front sandstones containing intercalations of marine siltstone and mudstone.

Beneath the Wasatch Plateau, strata of the Indianola Group pass laterally into time-equivalent beds of the Mancos Shale and Mesaverde Group exposed eastward on the Colorado Plateau (Weimer 1960; Lawton 1982). The Mesaverde Group is also a progradational clastic wedge that coarsens upward and intertongues eastward with the underlying Mancos Shale of marine origin (Fisher et al. 1960). Mesaverde beds include sandstones and siltstones of delta-front and delta-plain facies that pass upward and laterally into fluvial sequences dominated by sandstones, which are locally pebbly in the western part of the study area and in higher parts of the section farther east (Keighin and Fouch 1981; Fouch et al. 1983; Lawton 1983a, 1986).

Regional stratigraphic relations thus imply that the coarse clastics in the combined Indianola-Mesaverde progradational assemblage were derived mainly from the Sevier thrust belt lying to the west of the Cretaceous foreland basin (Spieker 1946). Only the upper part of the Mesaverde Group and the uppermost formation of the Indianola Group are treated in detail here. These units were deposited at a time when the migratory Sevier thrust system had advanced to a position at or near its maximum eastward extremity, where it provided a nearby source for clastic detritus shed into the adjacent foreland basin of central Utah.

Cretaceous-Tertiary Transition

Eroded frontal folds of the Sevier thrust system were overlapped unconformably by Paleogene strata of the North Horn Formation and overlying units (Stanley and Collinson 1979; Jefferson 1982). Folded synorogenic conglomerates equivalent to the Price River Formation within the upper Sixmile Canyon Formation (see Fig. 4) are structurally concordant with the remainder of the Indianola Group, whereas postorogenic conglomerates of the North Horn Formation are nearly flat-lying except where tilted within normal fault blocks of Neogene age (Lawton 1985).

The unconformity beneath the North Horn Formation extends across the full width of the study area, although its angularity is slight except near the thrust front, and its presence is commonly marked only by local pebbly lag deposits (Lawton 1986). In the central part of the Wasatch Plateau, North Horn sedimentation began during the Maastrichtian and continued without break into the Paleogene (Spieker 1946). The North Horn depocenter lay within a residual foreland trough that developed between the dormant Sevier thrust belt and the San Rafael uplift (Fig. 4), whose growth began before the end of Cretaceous time (Lawton 1985, 1986). North Horn lithofacies represent varied fluviatile environments together with locally developed paludal and lacustrine environments (Fouch 1975; Lawton 1986). Cretaceous horizons of the North Horn Formation are essentially confined to the troughlike depocenter, but Paleocene horizons onlap both the truncated thrust belt structures to the west and the eroded crest of the San Rafael uplift to the east (Fouch et al. 1983; Lawton 1986). The North Horn Formation reaches a maximum thickness of only about 500 m on the Wasatch Plateau within the study area (Birsa 1973), but time-equivalent conglomeratic fluvial strata of the Currant Creek Formation northwest of the Uinta Basin (see Fig. 2) are about 1,500 m thick (Isby and Picard 1983).

Paleogene Units

Conformable with the North Horn overlap assemblage is a thick overlying succession of nonmarine Paleogene deposits whose varied facies occur in complex intertonguing relationships (Ryder et al. 1976). These strata reach a maximum thickness of nearly 5,000 m within the depocenter of the Uinta Basin, a Laramide downwarp formed by structural disruption of the foreland region during latest Cretaceous and Paleogene time (Fouch 1975; Bruhn et al. 1983). The Paleogene sedimentary succession is overlain unconformably by Early Oligocene volcanics (32–33 m.y. BP) in the area where Paleogene strata overlap the dormant Sevier thrust belt (Jefferson 1982).

Depositional facies within the Paleogene sequence include varied lacustrine, deltaic, and fluvial deposits (Picard 1957b; Fouch 1975; Ryder et al. 1976; Moncure and Surdam 1980). The Uinta Basin was a lacustrine depocenter surrounded by tracts of deltaic and fluvial facies (Picard 1955). The central lacustrine facies is typified by dark shales of the Green River Formation. Inherently rapid facies changes (Picard and High 1972) between this central facies and various marginal facies were amplified and complicated by episodic fluctuations in lake level (Picard and High 1968). These fluctuations caused dramatic shifts in the positions of lithofacies boundaries and consequently produced markedly cyclic patterns of fluvial and lacustrine sedimentation. Lacustrine environments varied from deep water to playa conditions (Ryder et al. 1976).

The stratigraphic nomenclature developed for the Paleogene sequence is necessarily intricate (Fouch 1976), but need be reviewed only briefly here (see Fig. 4). Spieker (1946) regarded the North Horn Formation, Flagstaff Limestone, and Colton Formation of the Wasatch Plateau as formations of the Wasatch Group beneath the Green River Formation, and this formational framework is still used in the Sanpete Valley area (Stanley and Collinson 1979). Within the Uinta Basin, however, strata equivalent to the Flagstaff Limestone are referred to the Flagstaff Member of the Green River Formation (Fouch 1976). In that usage, the Green River Formation thus encloses an intertonguing body of the fluviodeltaic, sand-rich Colton Formation (Murany 1964; Cashion 1967). Some workers (Picard 1955, 1957a, 1967, 1971; Picard and High 1970) have subdivided the intertonguing lacustrine and related facies of the Green River Formation into the formal members of Bradley (1931), but others (e.g., Ryder et al. 1976) dispense with that nomenclature and instead use key marker horizons to establish the overall geometry of the various sedimentological facies present. The Green River Formation reaches thicknesses of 1,500 to 2,500 m in the subsurface within the Uinta Basin depocenter (Picard 1957a; Fouch 1975; Bruhn et al. 1983). The Colton Formation has a maximum thickness of about 1,000 m on the outcrop in Desolation Canyon of the Green River (Cashion 1967), but locally may be 1,250 to 1,500 m thick in the nearby subsurface to the northwest (Picard 1957a).

The lacustrine strata of the Green River Formation are overlain by about 500 m of fluvial beds in the Uinta Formation along a time-transgressive gradational contact (Dane 1954; Cashion 1967). Fluvial deposits above the contact on the east grade westward to lacustrine strata that are continuous with the underlying Green River Formation in the center of the Uinta Basin (Ryder et al. 1976). Along the northern flank of the Uinta Basin, the Uinta Formation is overlain by partly conglomeratic

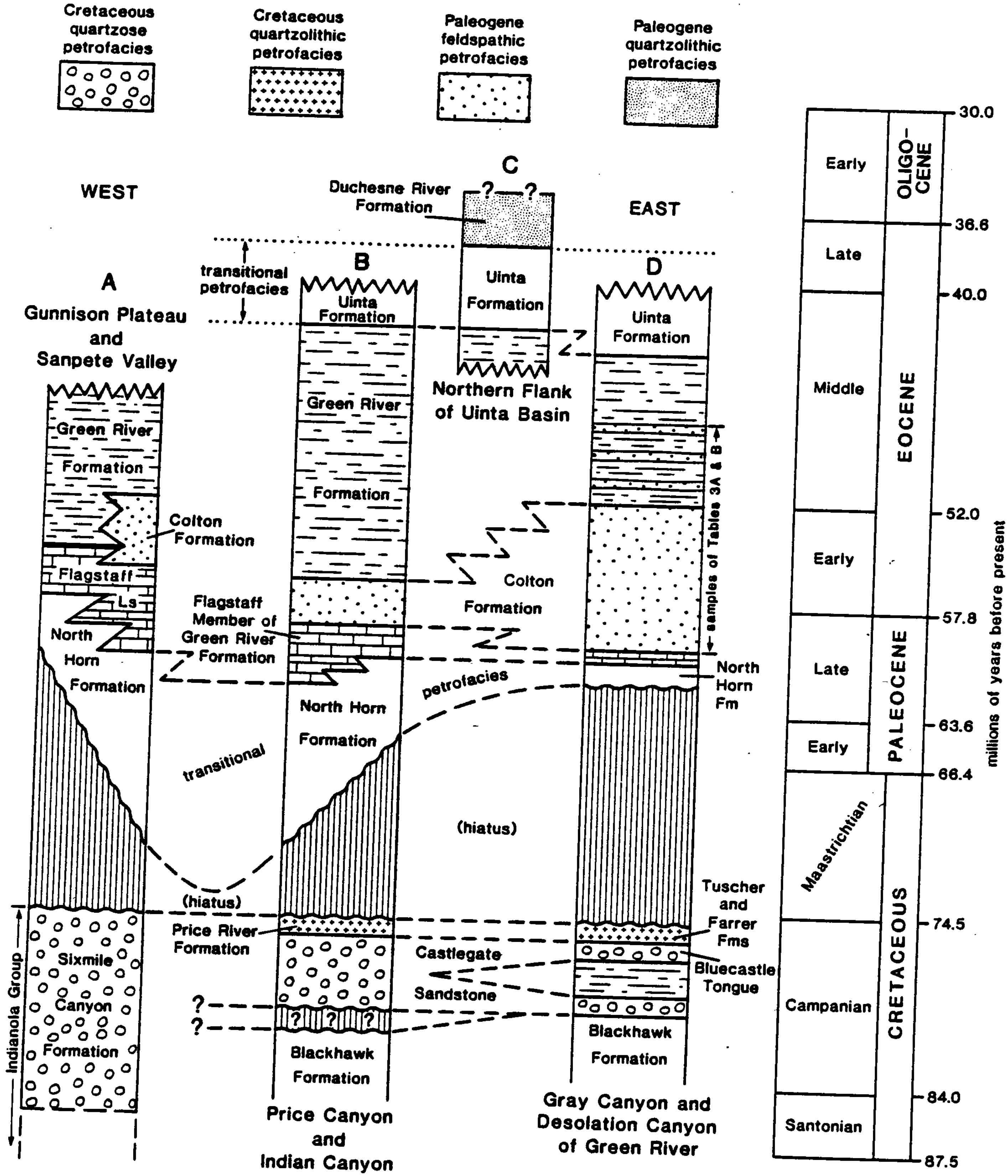


FIG. 5.—Schematic columnar sections showing stratigraphic distribution of key sandstone petrofacies within central Utah foreland (see Fig. 6 for locations of columns). Adapted after Dane (1954), Andersen and Picard (1972, 1974), Fouch (1975, 1976), Fouch et al. (1976, 1982, 1983), Ryder et al. (1976), Stanley and Collinson (1979), Zawiskie et al. (1982), and Lawton (1985, 1986). Time scale after Palmer (1983).

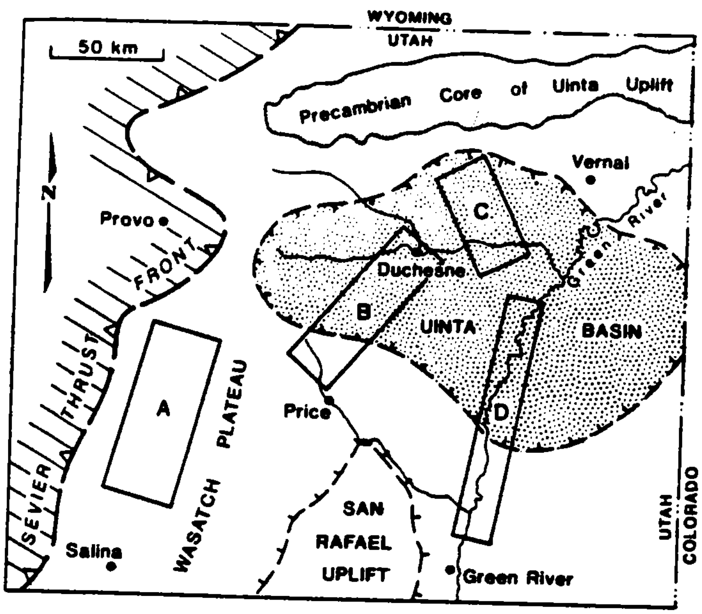


FIG. 6.—Sketch map of Uinta Basin and vicinity showing location of schematic columnar sections shown by Figure 5.

piedmont fluvial strata about 900 m thick in the Duchesne River Formation, whose basal contact is largely unconformable (Andersen and Picard 1974). Locally, however, the Uinta and Duchesne River Formations intertongue gradationally through a stratigraphic interval at least 300 m thick (Andersen and Picard 1972).

PETROFACIES DISTRIBUTION

Figure 5 denotes the overall stratigraphic distribution of key sandstone petrofacies within the study area at localities indicated by Figure 6. Two distinctive Cretaceous petrofacies, the Cretaceous quartzose petrofacies and the Cretaceous quartzolithic petrofacies, are present within the upper part of the Mesaverde Group deposited along the flank of the Cretaceous foreland basin, and the former is also present in equivalent strata of the Sixmile Canyon Formation at the top of the Indianola Group near the Sevier thrust front. Two additional Paleogene petrofacies, the Paleogene feldspathic petrofacies and the Paleogene quartzolithic petrofacies, occur within the sedimentary fill of the Laramide Uinta Basin. The former is typified by sandstones of the Colton Formation, whereas the latter is typified on the outcrop by younger sandstones of the Duchesne River Formation.

Representative detrital modes of the four key petrofacies are shown graphically by Figures 7 and 8 for Cretaceous and Paleogene sandstones, respectively. The most mature frameworks are those of the Cretaceous quartzose petrofacies, which plots near the Qm pole. Both Cretaceous and Paleogene quartzolithic petrofacies plot near the Qm-Lt leg in positions that commonly reflect recycling of supracrustal detritus. The Paleogene feldspathic petrofacies has an arkosic composition near the Qm-F leg in a position that commonly reflects derivation from plutonic or basement sources. Intervals of transitional petrofacies with variable compositions occur near the Cretaceous-Tertiary boundary at the base of the Paleogene overlap assemblage, and within Eocene strata in the interior of the Uinta Basin (see Fig. 5).

The following sections discuss the stratigraphic context and petrologic nature of each of the four key petrofacies in order of dominant age and include brief comments on the nature of the transitional petrofacies. The closing section of the paper is a discussion of the inferred Paleogene paleogeography of the study area and surrounding region based on previous work and the insights we have gained from our petrofacies analysis.

CRETACEOUS QUARTZOSE PETROFACIES

The Cretaceous quartzose petrofacies occurs in the Sixmile Canyon Formation of the Indianola Group along the Sevier thrust front, and in the Castlegate Sandstone, including the Bluecastle Tongue, of the upper Mesaverde Group east of the Wasatch Plateau (Fig. 5). Castlegate sedimentation began in mid-Campanian time shortly after 80 m.y. BP, and ended well before the end of the Campanian (Fouch et al. 1983; Lawton 1986). Deposition of the Sixmile Canyon Formation occupied a somewhat longer time span bracketed by palynomorphs of Santonian and uppermost Campanian age (Lawton 1985, 1986). Detrital modes of sandstones from the two units are statistically indistinguishable (Table 1).

Strata containing the Cretaceous quartzose petrofacies display easterly to southeasterly paleocurrent trends (Fig. 3), which suggest derivation of the quartz-rich sand from uplifted Precambrian and Paleozoic sedimentary rocks in the Sevier thrust belt to the west and northwest. Variable but locally high contents of detrital limeclasts (see Table 1) are compatible with such a sedimentary source terrane containing limestone as well as quartzite. Dominant clast types in conglomerates of the Indianola Group are quartzites, carbonate rocks, and chert derived from rock sequences now exposed within the thrust belt (Lawton 1983b). Chert clasts were derived mainly from replacement nodules present in the carbonate rocks.

Facies relations strengthen the paleocurrent evidence for a nearby provenance within the thrust belt. The Sixmile Canyon Formation, about 1,500 m thick, is composed of coarse piedmont clastics that prograded eastward during the time of emplacement of the Charleston-Nebo allochthon (Jefferson 1982), the most easterly major thrust sheet in the adjacent Sevier thrust belt. Farther east, sandy braidplain and coarse-grained meanderbelt deposits of the coeval Castlegate Sandstone, about 200 m thick at Price Canyon, formed a sandy coastal plain that prograded eastward across the top of an older deltaic complex represented by the underlying Blackhawk Formation. The sharp contact between the two units marks a change from meandering to braided fluvial deposition and probably represents a diastem rather than an erosional disconformity (Lawton 1986). Still farther east (Figs. 4–5), Castlegate fluvial sandstones intertongue conformably with finer-grained marine and marginal-marine strata well exposed along the Green River (van de Graaff 1972). The widespread dispersal of sheetlike fluviatile and strandline sand bodies within the Castlegate Sandstone

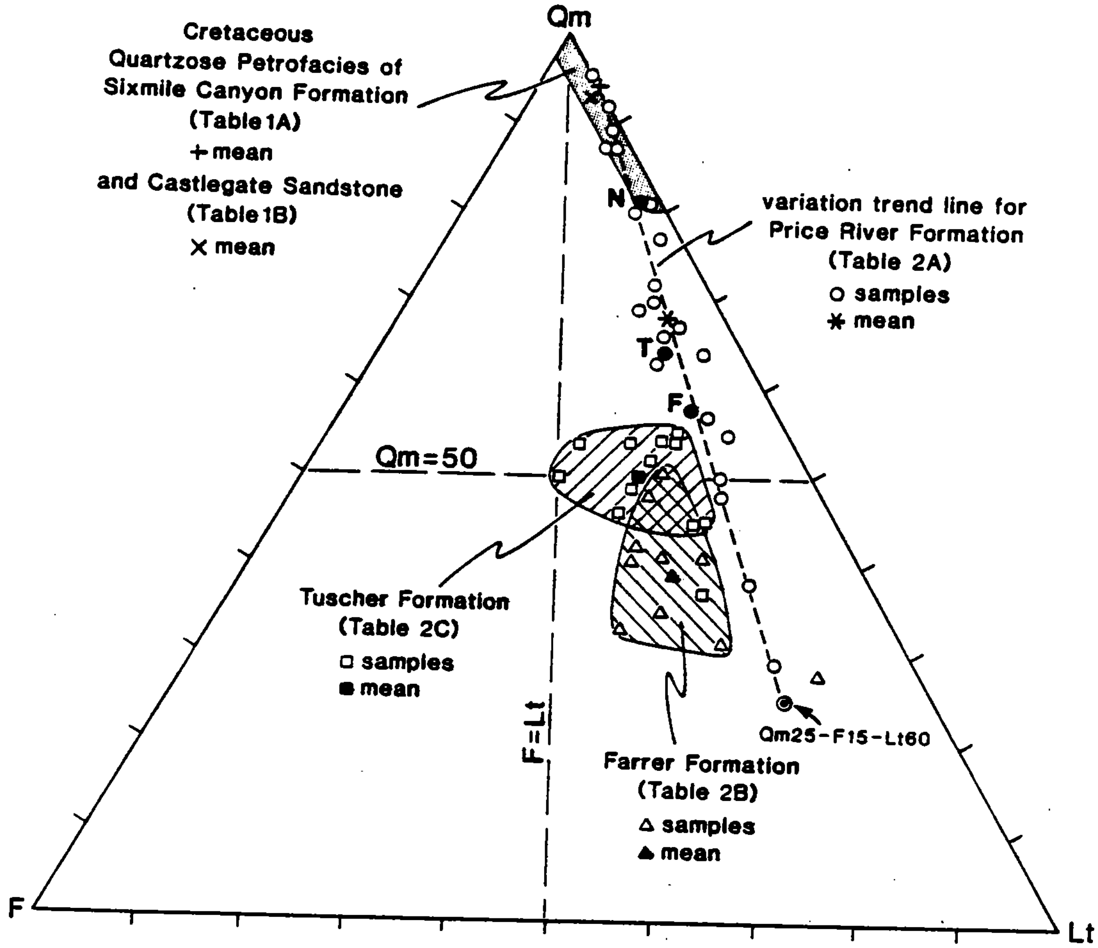


FIG. 7.—QmFLt plot (Lt = L + Qp) for detrital modes of sandstones from the Cretaceous quartzose petrofacies (shaded) and quartzolithic petrofacies (Price River, Farrer, and Tuscher Formations) in central Utah foreland (see Tables 1–2). Solid circles are formation means (after Keighin and Fouch 1981) for subsurface samples of the Neslen (N), Farrer (F), and Tuscher (T) Formations in the northeastern Uinta Basin (see text for discussion). Limeclasts (Lc) not included in Lt (see text).

may represent a sedimentary response to pulses of deformation within the thrustbelt provenance (Spieker 1946; van de Graaff 1972). Episodic intervals of sediment bypass across the piedmont belt may account for the likelihood that the Sixmile Canyon Formation is an incomplete section within which several hiatuses of uncertain magnitude are present (Fouch et al. 1982, 1983).

The low feldspar content of the Cretaceous quartzose petrofacies derived from the Sevier thrust belt contrasts with the compositions of sandstones in older Mesozoic sequences within the Utah foreland region. Available data suggest that the older Mesozoic sandstones of contrasting provenance have a mean composition of about Qt85-F11-L4 (Bruhn et al. 1983), and thus contain more feldspar by an order of magnitude (Table 1). The additional feldspar may have been derived in different instances either from basement rocks of the continental craton to the east or from igneous rocks in interior parts of the orogenic belt to the west.

CRETACEOUS QUARTZOLITHIC PETROFACIES

The Cretaceous quartzolithic petrofacies (Table 2) is composed of the uppermost Campanian (Fouch et al. 1983; Lawton 1986) Price River Formation, 190 m thick at Price Canyon, and the correlative Farrer (290 m) and Tuscher (280 m) Formations east of the San Rafael uplift near the Green River (Fig. 5). Paleocurrent trends are northeasterly, subparallel to the Sevier thrust belt (Fig. 3). Lateral continuity between the Price River Formation and its easterly equivalents was disrupted by erosion along the crest of the San Rafael uplift (Lawton 1986) prior to deposition of the North Horn Formation (Fig. 4).

The Price River Formation was deposited by a sinuous fluvial system flowing in front of the Sevier thrust belt (Lawton 1986). Price River quartzolithic sandstones plot as an elongate array that represents various admixtures of quartzose detritus, apparently identical in composition to that in the Cretaceous quartzose petrofacies, with lithic detritus having a mode near Qm25-F15-Lt60 (Fig. 7). We infer that this spectrum of quartzolithic compositions reflects the mingling of varied supracrustal detritus delivered to the foreland region along various segments of the Sevier thrust front, and then transported jointly within the longitudinal drainage system that prevailed during latest Campanian sedimentation. The feldspar content of the more lithic sandstones suggests that igneous rocks lying west of the Sevier thrust belt within the interior of the orogenic belt may have formed part of the provenance, but the content of sedimentary lithic fragments (Ls) consistently exceeds that of volcanic lithic fragments (Lv) and suggests that sources within the Sevier thrust belt were dominant (Lawton 1983b).

Correlative Farrer and Tuscher sedimentation farther

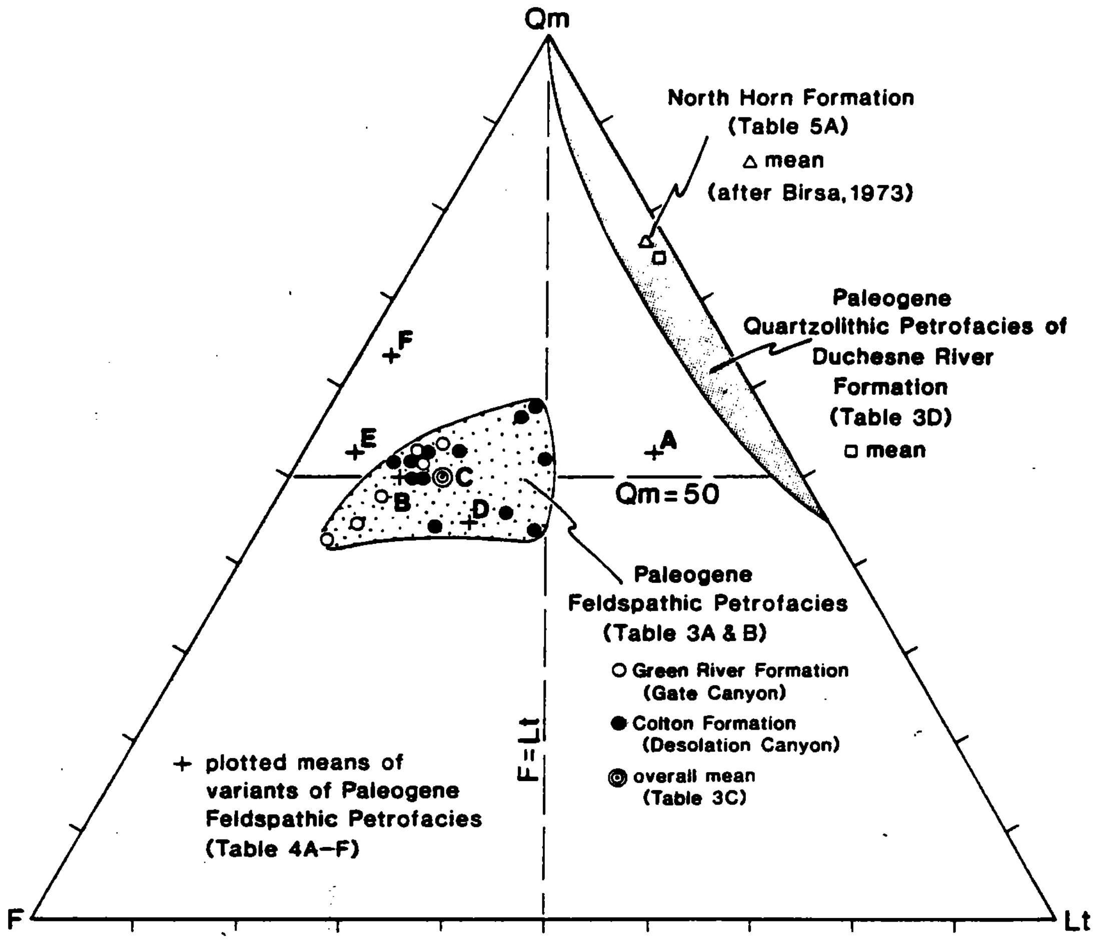


FIG. 8.—QmFLt plot (Lt = L + Qp) for detrital modes of sandstones from the Paleogene feldspathic petrofacies (stippled) and quartzolithic petrofacies (shaded) in the central Utah foreland (see Tables 3–5). Limeclasts (Lc) not included in Lt (see text).

east recorded the evolution of a fluvial system from meandering trunk streams and low sand/shale ratios during Farrer deposition to increasingly less sinuous channel patterns and high sand/shale ratios during Tuscher deposition, which closed with the development of a pebbly braidplain (Lawton 1986). Pebbly beds that rest disconformably on the Farrer Formation near Range Creek (Fig. 3), but lie unconformably beneath the North Horn Formation, are interpreted as local onlap of the Tuscher Formation along the eastern flank of the San Rafael uplift, whose initial (?) growth is thereby dated as latest Campanian (Lawton 1986).

Farrer and Tuscher quartzolithic sandstones are consistently (Fig. 7) and significantly (Table 2) more feldspathic than correlative sandstones in the Price River Formation, and some contain more volcanic (Lv) than sedimentary (Ls) lithic fragments (Lawton 1983b). As they were deposited by an integrated longitudinal drainage lying within the foreland region more than 100 km from the Sevier thrust front (Fig. 2), their provenance may have included igneous rocks within the interior of the orogenic belt, or to the south of the backarc thrust system in southern California and/or southern Arizona (Fig. 1). These possibilities cannot be tested readily because equivalent strata have been removed by erosion from the Henry Mountains region (Fig. 2) immediately to the southwest of the study area (Peterson and Ryder 1975; Peterson et al. 1980). Reportedly arkosic sandstones (Bowers 1972, p. B16) of the Kaiparowits Formation near the Kaiparowits Plateau still farther to the southwest (Fig. 2) are at least in part correlative with the Farrer-Tuscher sandstones (Peterson and Ryder 1975; Peterson et al. 1980) and may afford the means to trace the origin of the feldspathic detritus that is intermixed in the latter.

Keighin and Fouch (1981) have reported petrographic data for subsurface samples taken from the northeastern Uinta Basin (Fig. 2) at horizons correlative with the exposed Cretaceous quartzolithic petrofacies that we sampled on outcrop along the southwestern flank of the Uinta Basin. They studied 36 samples selected from a cumulative 36 m of core. Their detrital modes are similar to ours, except that quartz/feldspar ratios are consistently higher. For the Farrer Formation, they report a mean ($n = 9$) of Qt67-F8-L25 (vs. Qt54-F19-L27 from Table 2B), and for the Tuscher Formation, a mean ($n = 14$) of Qt70-F8-L22 (vs. Qt62-F17-L21 from Table 2C). Their samples from the underlying Neslen Formation have a mean ($n = 13$) of Qt84-F2-L14, comparable to our mean of Qt78-F5-L17 for the Price River Formation (Table 2A). Moreover, all three of their formation means plot along the spectrum of compositions for the Price River Formation on the QmFLt plot (Fig. 7), and clearly represent local variants of the Cretaceous quartzolithic petrofacies.

TABLE 1.—*Mean detrital modes[1] of selected sandstones from the Cretaceous quartzose petrofacies[2] in Central Utah (see text for symbols and discussion of grain types)*

Formation	A Sixmile[3]	B Castlegate[4]	C Combined
N	10	20	30
Qm	93.7 ± 5.1 (Ra 82–99)	93.1 ± 4.9 (Ra 81–99)	93.4 ± 5.0
Qp	5.0 ± 4.4 (Ra 0–14)	4.5 ± 3.8 (Ra 0–15)	4.6 ± 4.0
F	0.1 ± 0.3 (Ra 0–1)	0.8 ± 1.1 (Ra 0–4)	0.5 ± 1.0
L	1.2 ± 1.3 (Ra 0–4)	1.6 ± 1.7 (Ra 0–7)	1.4 ± 1.6
Summation	100	100	100
Lc[5]	12.6 ± 15.0 (Ra 0–40)	5.4 ± 10.1 (Ra 0–46)	7.8 ± 12.4

[1] Grain percentages ± standard deviations; total observed ranges in parentheses.
[2] See Figures 3–5 for geographic location and stratigraphic context of sampled horizons.
[3] Upper Cretaceous (Santonian–Campanian) Sixmile Canyon Formation of Indianola Group at Sixmile Canyon on west side of Wasatch Plateau at flank of Sanpete Valley; most proximal part of Cretaceous quartzose petrofacies of foreland basin near structural front of Sevier thrust belt.
[4] Upper Cretaceous (middle–upper Campanian) Castlegate Sandstone (including Bluecastle Tongue) of Mesaverde Group in Price Canyon and in Gray Canyon of the Green River; more distal part of Cretaceous quartzose petrofacies of foreland basin along line of Book Cliffs.
[5] Percentage of total framework grain population.

TABLE 2.—*Mean detrital modes[1] of selected sandstones from the Cretaceous quartzolithic petrofacies[2] in Central Utah (see text for symbols and discussion of grain types)*

Formation	A Price River[3]	B Farrer[4]	C Tuscher[4]
N	21	10	12
Qm	68.4 ± 17.4 (Ra 29–95)	39.1 ± 6.9 (Ra 28–51)	49.6 ± 5.3 (Ra 37–55)
Qp	9.4 ± 4.0 (Ra 2–20)	14.8 ± 4.6 (Ra 7–22)	12.8 ± 5.1 (Ra 5–20)
F	5.4 ± 3.9 (Ra 0–14)	19.0 ± 4.3 (Ra 10–27)	16.7 ± 4.1 (Ra 11–25)
L	16.8 ± 11.5 (Ra 0–44)	27.1 ± 9.5 (Ra 15–45)	20.9 ± 4.4 (Ra 13–27)
Summation	100	100	100
Lc[5]	3.5 ± 4.3 (Ra 0–14)	5.1 ± 3.6 (Ra 0–13)	1.2 ± 2.2 (Ra 0–8)

[1] Grain percentages ± standard deviations; total observed ranges in parentheses.
[2] See Figures 3–5 for geographic location and stratigraphic context of sampled horizons.
[3] Upper Cretaceous (upper Campanian) Price River Formation of Mesaverde Group at Price Canyon; western subfacies of Cretaceous quartzolithic petrofacies in foreland basin between structural front of Sevier thrust belt and crest of San Rafael uplift.
[4] Upper Cretaceous (upper Campanian) Farrer and Tuscher Formations of Mesaverde Group at Range Creek and in Gray Canyon of the Green River; lower (Farrer) and upper (Tuscher) parts of eastern subfacies of Cretaceous quartzolithic petrofacies in foreland basin east of crest of San Rafael uplift.
[5] Percentage of total framework grain population.

Quartzolithic sandstones were also deposited within the unconformably overlying North Horn Formation (Stanley and Collinson 1979) following a prominent hiatus of as much as 10–15 m.y. during the Maastrichtian and Paleocene (Fig. 5). However, Paleocene North Horn sandstones also include more feldspathic types transitional to the Paleogene feldspathic petrofacies. Consequently, we provisionally assign the North Horn Formation to a transitional petrofacies composed in part of locally reworked detritus.

PALEOGENE FELDSPATHIC PETROFACIES

The Paleogene feldspathic petrofacies occurs in fluvial and deltaic sandstone bodies of the Colton Formation, and in sandstone tongues within the overlying Green River Formation of dominantly lacustrine origin (Fig. 5). The Colton and North Horn Formations are separated by an interval of lacustrine beds in the Flagstaff Member of the Green River Formation or Flagstaff Limestone of Late Paleocene to Early Eocene age (Fouch 1976; Stanley and Collinson 1979). The Colton Formation is also dominantly of Late Paleocene and Early Eocene age (Fouch 1976; Ryder et al. 1976) but may include strata as young as Middle Eocene near Sanpete Valley (Zawiskie et al. 1982). The overlying Green River Formation of Early to Middle Eocene age contains tuff beds that have been dated radiometrically at about 45 m.y. BP (Pitman et al. 1982).

Intertonguing with lacustrine shales of the Green River Formation causes the thickness and age of the Colton Formation to vary widely within the study area (Figs. 4–5). Its outcrop thickness increases eastward from the Wasatch Plateau (Zawiskie et al. 1982) and reaches a maximum of about 1,000 m in Desolation Canyon of the Green River (Picard 1955; Cashion 1967), where the sand-rich Colton strata conformably overlie about 25 m of lacustrine limestone and fine-grained clastic strata in the Flagstaff Member of the Green River Formation. The proportion of sandstone in the Colton Formation also increases eastward on the outcrop (Chapman 1982) from Price Canyon to the Green River (Figs. 3–4), and the sand/shale ratio within the intertonguing Green River and Colton Formations in the subsurface of the Uinta Basin similarly increases to the southeast toward the exposures in Desolation Canyon (Murany 1964). Paleocurrent trends toward the north and northwest for the Paleogene feldspathic petrofacies (Fig. 3) imply that the thick section of Colton Formation in Desolation Canyon was deposited near a major entry point for arkosic detritus moving into the Uinta Basin.

Detrital modes used to define the composition of the Paleogene feldspathic petrofacies (Table 3A, B) were determined for 18 samples collected from horizons spaced 30 to 150 m apart within a composite section 1,200 m thick including the entire Colton Formation and the lower part of the overlying Green River Formation up to the "Horse Bench Sandstone Bed" of Middle Eocene age (Fig. 4). The Colton Formation was sampled along Desolation Canyon (accessible only by boat), and the Green River

TABLE 3.—*Mean detritial modes[1] of selected sandstones from Paleogene feldspathic (A–C) and quartzolithic (D) petrofacies in the Uinta Basin (see text for symbols and discussion of grain types)*

Formation	A Colton[3]	B Green River[4]	C A and B	D Duchesne River[5]
N	12	6	18	46
Qm	50.9 ± 4.3 (Ra 44–58)	49.2 ± 4.1 (Ra 43–54)	50.4 ± 4.3 (Ra 43–58)	75.2 ± 12.0 (Ra 47–98)
Qp	3.0 ± 2.0 (Ra 0–6)	1.3 ± 0.7 (Ra 0–2)	2.4 ± 1.9 (Ra 0–6)	11.0 ± 6.4 (Ra 0–35)
F	32.3 ± 6.0 (Ra 22–39)	40.5 ± 6.0 (Ra 33–50)	35.0 ± 7.2 (Ra 22–50)	1.4 ± 1.4 (Ra 0–6)
L	13.8 ± 4.5 (Ra 9–21)	9.0 ± 1.6 (Ra 6–11)	12.2 ± 4.4 (Ra 6–21)	12.4 ± 11.8 (Ra 0–45)
Summation	100	100	100	100
Lc[6]	5.3 ± 5.1 (Ra 0–17)	1.3 ± 1.4 (Ra 0–4)	4.0 ± 4.6 (Ra 0–17)	3.0 ± 3.4 (Ra 0–15)

[1] Grain percentages ± standard deviations; total observed ranges in parentheses.
[2] See Figures 3–5 for geographic location and stratigraphic context of sampled horizons.
[3] Upper Paleocene to Lower Eocene Colton Formation of the Roan Cliffs in Desolation Canyon (Fig. 9) of the Green River; lower part of Paleogene feldspathic petrofacies along southern flank of Uinta Basin.
[4] Lower to Middle Eocene Green River Formation up to "Horse Bench Sandstone Bed" in Gate Canyon (Fig. 9); upper part of Paleogene feldspathic petrofacies along southern flank of Uinta Basin.
[5] Upper Eocene Duchesne River Formation in drainages north of Duchesne River (Andersen and Picard 1972, 1974); upper part of Paleogene quartzolithic petrofacies along northern flank of Uinta Basin; tabulated from data of D. W. Andersen.
[6] Percentage of total framework grain population.

Formation was sampled in nearby Gate Canyon (Fig. 9). Approximately 100 newly recorded paleocurrent indicators (mostly trough-set crossbeds) were measured along traverses up eight side canyons tributary to Desolation Canyon (Fig. 9) in order to confirm net sediment dispersal toward the north-northwest.

A mean composition of Qm50-F35-Lt15 (Table 3C) suggests derivation of the Paleogene feldspathic petrofacies (Fig. 8) from dominantly plutonic or basement sources, as Picard (1971) indicated from his work east of the Green River. This inference is supported by the dominance of common plutonic quartz grains, the abundance of grid-twinned microcline in the potassium feldspar population, the occurrence of coarse muscovite and biotite as accessory minerals, and the presence of microgranitic rock fragments in some samples. Plagioclase and potassium feldspar are present in roughly subequal amounts (mean P/K ratio is 1.25). Subordinate lithic fragments are mainly quartzose metasedimentary (Ls) and felsic to intermediate volcanic (Lv) types.

Mean compositions determined by others for variants of the Paleogene feldspathic petrofacies elsewhere within the Uinta Basin are compared to our data in Table 4 and plotted with our data on Figure 8. Suites from west of Desolation Canyon are comparable (Table 4B) or less feldspathic (Table 4A). The latter composition may reflect dilution of arkosic detritus with quartzolithic debris reworked from the vicinity of the Sevier thrust belt. Suites

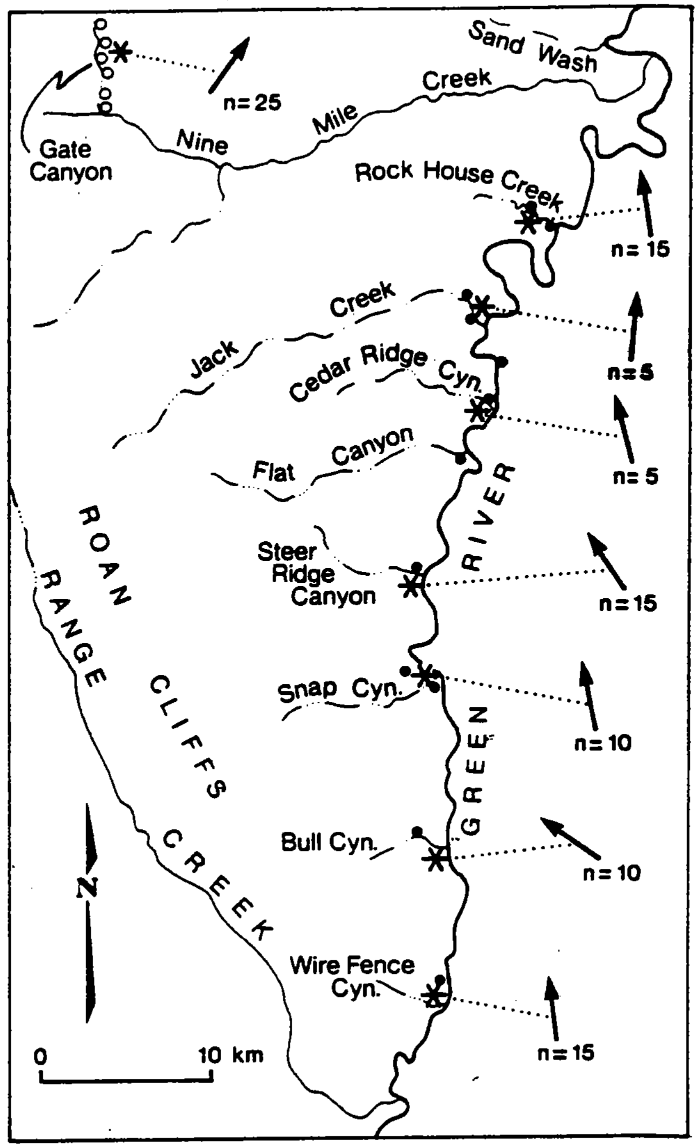


FIG. 9.—Sketch map of Desolation Canyon (Sand Wash to Wire Fence Canyon) of the Green River and vicinity (see Fig. 3 for location) showing a) sample localities for 18 arkosic sandstones (summary of point counts in Table 2) from the Colton Formation (solid circles) and the lower part of the overlying Green River Formation (open circles) up to the "Horse Bench Sandstone Bed" (see Figs. 4–5 for stratigraphic position); and b) mean paleocurrent vectors (heavy arrows) at 8 traverse sites (asterisks).

from east of Desolation Canyon (Table 4E, F) are seemingly less lithic, but we note that these suites reportedly contain an average of 20–25 percent matrix and cement (Picard 1971), whereas our counts averaged only 5–10 percent interstitial materials; conceivably, the differences in the calculated detrital modes may represent only operator variance in the treatment of lithic detritus. The more quartzose lacustrine sandstones (Table 4F) contain minor proportions of carbonate intraclasts and/or ooliths that are absent in the associated fluvial sandstones (Table 4E).

Pitman et al. (1982) studied a suite of sandstone samples selected from 15 m of subsurface core taken from

TABLE 4.—*Comparative mean detrital modes for selected sandstone suites of the Paleogene feldspathic petrofacies within the Uinta Basin*[1] *(UB)*

Suite	Location	Age	Unit(s)	N	Qm	F	Lt	Reference
A	Price Canyon and Soldier Creek, Southwest UB	Upper Paleocene and Lower Eocene	Colton Formation	15	53	13	34	Chapman 1982
B	Sunnyside Area, Southwest UB	Upper Paleocene and Lower Eocene	Colton Formation	25	50	39	11	Banks 1981
C	Gate Canyon and Desolation Canyon, Southern UB	Upper Paleocene to Middle Eocene	Colton and Green River Formations	18	50	35	15	Table 3
D	Pariette Bench Field (15-m core), Central UB	Lower Eocene	Green River Formation	20	45	35	20	Pitman et al. 1982
E	P. R. Spring Area (fluvial ss) Southeast UB	Lower Eocene to Middle Eocene	Colton and Green River Formations	19	53	42	5	Picard 1971
F	P. R. Spring Area (lacustrine ss) Southeast UB	Lower Eocene to Middle Eocene	Colton and Green River Formations	20	64	33	3	Picard 1971

[1] See Figures 2–5 and 10 for regional stratigraphic and tectonic setting of sample suites.

the lower part of the Green River Formation near the center of the Uinta Basin. They determined a mean composition (n = 20) of Qm45-F35-Lt20 (Table 4D), quite close to our mean value of Qm50-F35-Lt15 (Table 4C). As their method of collection can be regarded as effectively a spot sample of the subsurface sandstones, we conclude that the mean composition we have determined for the Paleogene feldspathic petrofacies is a valid predictor of compositions to be expected in subsurface equivalents.

We concur with previous workers who have inferred that the arkosic detritus in the Paleogene feldspathic petrofacies was transported into the region of the Uinta Basin from Laramide uplifts lying generally to the southeast (Picard 1957a, 1971; Stanley and Collinson 1979; Banks 1981; Chapman 1982; Pitman et al. 1982; Zawiskie et al. 1982; Bruhn et al. 1983). The appearance of the arkosic debris in the foreland region thus marked the transition from Sevier to Laramide tectonics. The paleotectonic and paleogeographic implications of this conclusion are discussed further in later sections.

The Uinta Formation of Middle to Late Eocene age (Fig. 5) includes feldspar-rich sandstones with a mean composition (n = 21) of Qm20-F68-Lt12 (Bruhn et al. 1983), even more feldspathic in composition than the Paleogene feldspathic petrofacies. However, the Uinta Formation locally intertongues with overlying strata of the Duchesne River Formation (see above), which contains sandstones of the Paleogene quartzolithic petrofacies. Consequently, the mixed and interbedded feldspathic and lithic sandstones (Andersen and Picard 1972, p. 9) of the Uinta Formation are provisionally assigned here to a transitional petrofacies.

PALEOGENE QUARTZOLITHIC PETROFACIES

The Paleogene quartzolithic petrofacies (Table 3D) occurs in the Duchesne River Formation of Late Eocene to Early Oligocene(?) age (Andersen and Picard 1972, 1974) along the northern flank of the Uinta Basin (Fig. 5). These strata form a fluviatile piedmont facies of proximal braidplain deposits and more distal floodplain deposits along the southern flank of the Uinta uplift. The most abundant lithic fragments in sandstones (Fig. 8) are quartzite, chert, clastic sedimentary types (Ls), and detrital limeclasts (Lc). Facies relations and consistently southerly paleocurrent trends (Fig. 3) leave no doubt that the sources of Duchesne River detritus were Precambrian sedimentary rocks in the core of the Uinta uplift and the Paleozoic sedimentary cover that was stripped from the core by erosion during Laramide time.

Similar quartzolithic sandstones with detrital modes that plot near the Qm-Lt leg (Isby and Picard 1983, p. 99) occur farther west in the Currant Creek Formation at the northwestern edge of the Uinta Basin (Fig. 2). These older strata, correlative with the North Horn Formation (see above), display southerly paleocurrent indicators (Fig. 10), but become generally finer-grained from west to east (Isby and Picard 1983). Compositionally similar quartzolithic sandstones (Table 5A) in the North Horn Formation (Fig. 8) were probably derived chiefly from the dormant Sevier thrust belt (Birsa 1973), rather than from the Uinta uplift. The quartzolithic sandstones in the Currant Creek Formation were deposited near the structural juncture between the Sevier thrust belt and the Uinta uplift, and may conceivably have been derived in part from both potential provenances. The evident similarity of detrital modes for sandstones derived from the Sevier thrust belt (Table 5A) and from the Uinta uplift (Table 5B) demonstrates that a) extensive knowledge of paleocurrents and facies patterns is required to distinguish the two as potential sources of detritus, and b) the Uinta uplift was definitely not a source of arkosic detritus, despite the abundance of feldspathic sandstones in the adjacent Uinta Basin (see above).

Some mid-Eocene sandstones intercalated within the Green River Formation in northern parts of the Uinta Basin also have quartzolithic frameworks (Table 5C, D).

TABLE 5.—*Comparative mean detrital modes for selected sandstone suites of the Paleogene quartzolithic petrofacies in and near the Uinta Basin*[1]

Suite	Location	Age	Unit(s)	N	Qm	F	Lt	Reference
A	Sanpete Valley, southwest of Uinta Basin	Maastrichtian to Paleocene	North Horn Formation	?	77	2	21	Birsa 1973
B	Northern Flank of Uinta Basin	Upper Eocene to Oligocene (?)	Duchesne River Formation	46	75	2	23	Table 3
C	Raven Ridge (fluvial ss), NE Uinta Basin	mid-Eocene (?)	Green River Formation	6	72	12	16	Picard and High 1972
D	Raven Ridge (lacustrine ss), NE Uinta Basin	mid-Eocene (?)	Green River Formation	15	89	5	6	Picard and High 1972

[1] See Figures 2–5 and 10 for regional stratigraphic and tectonic setting of sample suites.

These sandstones display southerly paleocurrents (Fig. 10), and contain detritus derived mainly from the Uinta uplift (Picard and High 1972), rather than from the other Laramide uplifts that fed arkosic debris to the southern flank of the Uinta Basin (see above). According to P. C. van de Kamp (cited as oral communicant by Ryder et al. 1976), subsurface samples indicate that the delivery of Uinta-derived quartzolithic sand to the northern flank of the Uinta Basin and feldspathic sand of unknown provenance to the southern flank of the Uinta Basin was a persistent pattern of dispersal throughout much of the Paleogene. In the subsurface of the northern Uinta Basin, therefore, quartzolithic sandstones are probably dominant in the Wasatch Formation, which is equivalent to both the North Horn and Colton Formations in areas where the intervening Flagstaff Member of the Green River Formation pinches out (Fouch 1975, 1976).

TECTONIC TRANSITION

Our petrofacies analysis fully supports the contention of Spieker (1949) that the deposition of Upper Cretaceous clastic wedges in the central Utah foreland was contemporaneous with deformation in the Sevier thrust belt, where orogenic highlands were the provenance for the bulk of the clastic sediment. Part of the sedimentary record for the regional transition from Sevier to Laramide tectonics is the change from quartz-rich petrofacies composed dominantly of recycled sedimentary detritus to Paleogene arkosic petrofacies in sandstones of the central Utah foreland. In this section, we review the evidence for the timing of the tectonic transition; the location of the arkosic provenance is discussed in the succeeding section.

Thrusting in the adjacent segment of the Sevier thrust belt was complete by latest Campanian or earliest Maastrichtian time, when the youngest conglomeratic piedmont clastics of the Indianola Group were deposited and folded during or immediately following deposition of the synorogenic Price River Formation farther east (Lawton 1985). Eroded frontal folds of the Sevier thrust belt were subsequently overlapped by postorogenic terrestrial facies of the North Horn Formation during the Paleocene (Fig. 5).

Evidence for the initial timing of deep-seated Laramide deformation east of the thin-skinned Sevier thrust belt is provided by stratigraphic relations across the San Rafael uplift (Fig. 4). Uppermost Campanian strata were thinned by erosion along the crest of the San Rafael uplift prior to onlap deposition of Paleocene beds in the North Horn Formation above a regional unconformity. There is thus little question that the San Rafael uplift was in existence by Maastrichtian time (e.g., Fouch et al. 1982, 1983), and the other western monoclines of the Colorado Plateau (Kelley 1955a, b) may have begun to form by that time also. The occurrence of pebble types that may represent recycled intrabasinal detritus within the uppermost Campanian Tuscher Formation suggests that tectonism may have begun to break up the integrated Cretaceous foreland basin in the region of the San Rafael uplift by latest Campanian time (Lawton 1986). This inference is supported by the presence of local pebbly lag deposits correlated provisionally with the Tuscher Formation along the eastern flank of the San Rafael uplift (see above). The intra-Campanian change in foreland paleocurrent trends from east-southeasterly for the Cretaceous quartzose petrofacies to northeasterly for the Cretaceous quartzolithic petrofacies can also be interpreted as evidence for the development of a northeasterly structural and topographic grain within the foreland region (see Fig. 3).

Convergent evidence from both the Sevier thrust front and the San Rafael uplift thus indicates that tectonism began to shift eastward into the region of the previously subsiding foreland basin in latest Campanian time and had been achieved by earliest Maastrichtian time. Southerly paleocurrents in the Currant Creek Formation suggest that the Uinta uplift may have begun its growth in Maastrichtian time (Bruhn et al. 1983), although most of its structural relief did not develop until Late Paleocene and Early Eocene time (Ryder et al. 1976). The Flagstaff lacustrine basin (Fig. 10) of Late Paleocene and Early Eocene age occupied a residual foreland trough that was trapped between the dormant Sevier thrust belt on the west and the uparched San Rafael uplift on the east (Stanley and Collinson 1979). The Uinta Basin continued to subside throughout the Eocene, and Lake Uinta reached its maximum extent during mid-Eocene time (Picard 1955; Ryder et al. 1976; Bruhn et al. 1983).

Although the Sevier-Laramide tectonic transition thus apparently occurred locally within the central Utah foreland before the end of Cretaceous time, arkosic sand reflecting a dominantly Laramide provenance did not reach the Uinta Basin from the southeast prior to mid-Paleocene time. Moreover, the arkosic detritus did not reach the region of the dormant thrust front near the Sanpete Valley until even later, in Early Eocene time (Fig. 5); its arrival there was probably delayed until burial of the San Rafael uplift during the Paleocene allowed sediment transport across the site of that structure. These observations imply that the major uplifts of the eastern Laramide province (Kelley 1955a, b) grew principally in Paleogene time, as structural studies suggest (Berg 1962), and persisted later than the Laramide uplifts of the Colorado Plateau region. Convincing evidence for analogous eastward migration of Laramide tectonism from Late Cretaceous to Eocene time has been assembled recently for southwest Montana (Schmidt and O'Neill 1982; Schmidt and Garihan 1983).

PALEOGENE PALEOGEOGRAPHY

Despite clearcut paleocurrent data (Figs. 3, 10) and unequivocal facies relations (see above) indicating that feldspathic detritus entered the region of the Uinta Basin from the south-southeast during the Paleogene, its ultimate source has never been clear. The problem has been to identify a potential source from which thousands of cubic kilometers of arkosic debris could have been eroded during the Paleogene.

Most of the monoclinal Laramide uplifts of the Colorado Plateau are largely mantled even today by Mesozoic sequences (Figs. 2, 10), which contain mostly quartz-rich sandstones and could not have yielded voluminous arkosic detritus during the Paleogene. The nearby Uncompahgre uplift fails as a significant potential provenance (Picard 1971) on this score alone. Stanley and Collinson (1979) suggested that mica-bearing feldspar-rich sandstones of the Pennsylvanian-Permian Cutler Formation (Mack 1978) in the Monument uplift (Figs. 2, 10) may have contributed some of the required detritus. However, the area of this inferred provenance and the thickness of the presumed source rocks within it are inadequate to account for the volume of arkosic sand found in the region of the Uinta Basin. The primary arkosic sources were almost certainly farther east in basement-cored uplifts, to which previous workers have also called general attention (e.g., Stanley and Collinson 1979; Banks 1981; Bruhn et al. 1983).

The Needle Mountains uplift (Figs. 2, 10) in the San Juan region of southwestern Colorado has been cited as a suitable provenance for the mixture of quartzofeldspathic mineral grains and combined metasedimentary (Ls) and volcanic (Lv) lithic fragments observed in the Paleogene feldspathic petrofacies of central Utah (Pitman et al. 1982; Chapman 1982; Zawiskie et al. 1982). The present area of the Needle Mountains uplift is far too limited to serve as the primary source for large quantities of sediment. However, the Laramide Needle Mountains uplift was but one projection of the huge San Luis uplift (Tweto 1975, 1980), most of which is now buried beneath Oligocene volcanic cover in the San Juan Mountains and Neogene basin fill within San Luis Valley along the Rio Grande Rift (Fig. 10). The internal morphology of the Laramide San Luis uplift was doubtless complex in detail, for buried Eocene sediments are known locally from the subsurface within its general outline (Chapin and Cather 1981, 1983). During the Paleogene, however, the present Needle Mountains, Gunnison, Sawatch, and Sangre de Cristo uplifts were linked together within the overall San Luis uplift as a positive block of exposed basement nearly as extensive as the Front Range and related uplifts farther north.

We propose that the main dispersal path for arkosic detritus found within the Paleogene feldspathic petrofacies of the Uinta Basin was a major river system with its source in the San Luis uplift (Fig. 10). The inferred paleodrainage flowed longitudinally along the southwest flank of the Uncompahgre uplift, passing to the east of the Monument uplift, and debouched into the Uinta Basin near present Desolation Canyon of the Green River. Bruhn et al. (1983) noted that "prodigious quantities" of sand were supplied to Lake Uinta through a large Paleogene delta system built in that area. Lake Flagstaff farther west was also filled during the Paleogene by the northwestward progradation of feldspathic clastics (Stanley and Collinson 1979), which we conclude moved across the then-buried San Rafael uplift (Fig. 10). Some of the Paleogene arkosic detritus carried into the region of the Uinta Basin may have been gathered from small tributary drainages tapping suitable monoclinal uplifts, most notably the Monument uplift, within the Colorado Plateau (see above). However, we infer that most of the Paleogene feldspathic petrofacies was derived directly from the voluminous basement mass of the San Luis uplift. Recycled debris of sedimentary parentage in the Paleogene quartzolithic petrofacies was simultaneously (Green River Formation) and subsequently (Duchesne River Formation) delivered to the northern flank of the Uinta Basin from sources in the rising Uinta uplift (Fig. 10).

Our hypothesis that the San Luis uplift was the major provenance for Paleogene arkosic sand in the central Utah foreland is strengthened by the fact that no other basement-cored Laramide uplift was in a position to deliver sediment directly to that region. The central Wyoming uplifts lay beyond the extensive Green River and Washakie Basins to the north of the Uinta uplift (Fig. 2). The Front Range and related uplifts were screened by the White River uplift with its sedimentary cover to the east of the Piceance Basin of Colorado. Debris from potential sources in the orogenic region of southern Arizona was diverted along an east-west paleodrainage within the Baca Basin (Cather and Johnson 1984) lying south of Black Mesa and the Defiance uplift (Fig. 2). Essentially all other drainages southeast of the Four Corners Platform must have flowed into the depocenter of the San Juan Basin (Fig. 10), and much of the detritus shed from the southwestern flank of the San Luis uplift doubtless entered the San Juan Basin as well. Sediment transported northward (Young and McKee 1978) across the Grand Canyon region (Fig. 2) into southwestern Utah was probably deposited southwest of the Circle Cliffs uplift (Fig. 10) in the Claron Formation (Anderson and Rowley 1975), which

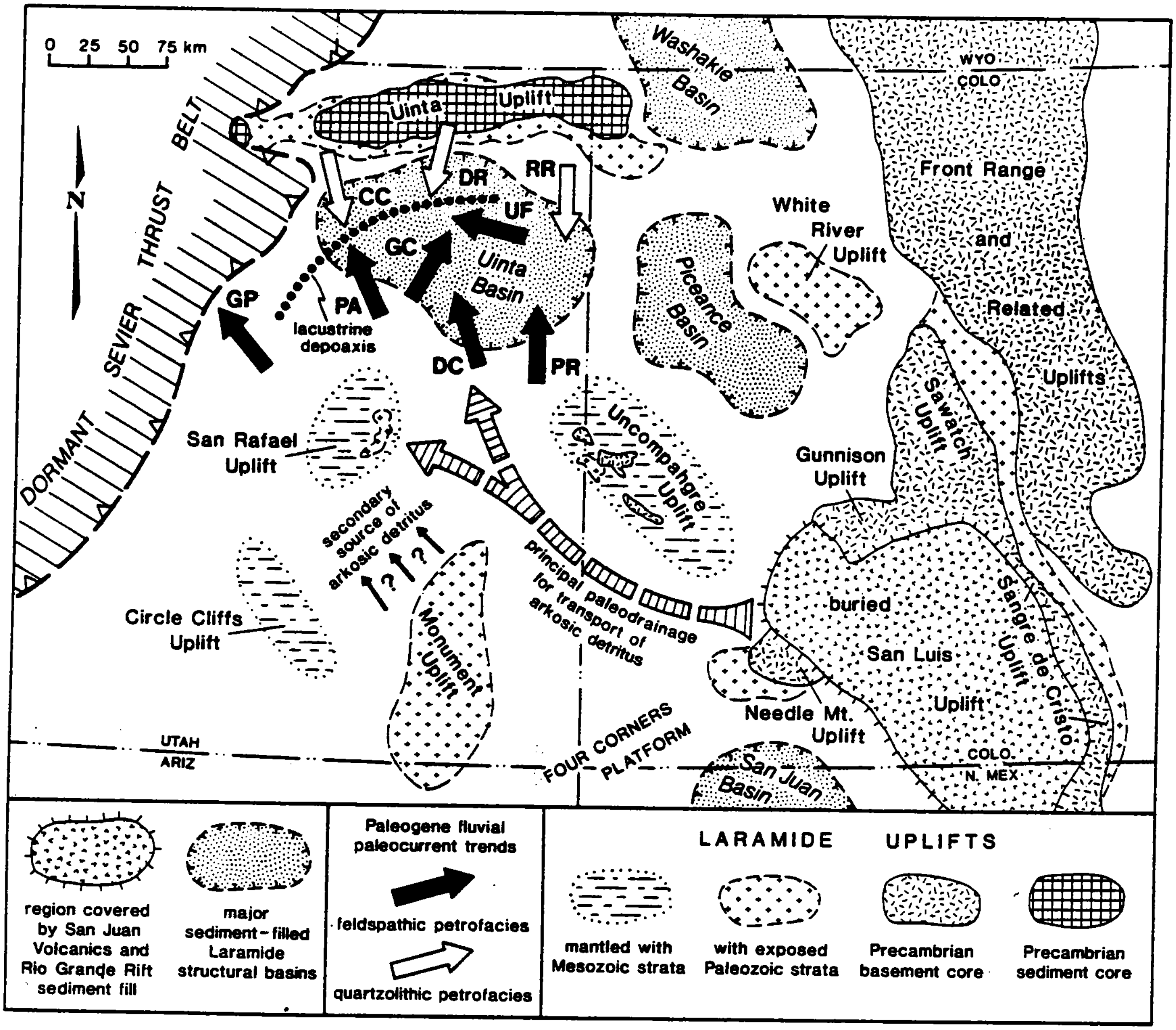


FIG. 10.—Inferred mid-Paleocene to mid-Eocene dispersal path of arkosic detritus from Laramide foreland uplifts of southern Rocky Mountains (Tweto 1975, 1980) and Colorado Plateau (Kelley 1955a, 1955b) into region of Uinta Basin. Paleogeography modified after Picard (1971), Stanley and Collinson (1979), Pitman et al. (1982), Zawiskie et al. (1982), Chapman (1982), and Bruhn et al. (1983). Lacustrine depoaxis refers to lacustrine environments of Lake Flagstaff (NE-SW trend on SW) and Lake Uinta (E-W trend on NE) after Ryder et al. (1976), Stanley and Collinson (1979), and Bruhn et al. (1983). Indicated paleocurrent localities within and near Uinta Basin (in mid-Paleocene to mid-Eocene strata except as noted): CC, Currant Creek Formation (Cretaceous to Paleocene) after Isby and Picard (1983); DC, Desolation Canyon of the Green River (Figs. 3, 9); DR, Duchesne River Formation (Upper Eocene) after Andersen and Picard (1972, 1974 and Fig. 3); GC, Gate Canyon (Figs. 3, 9) and Sunnyside area (Fig. 3 after Chapman 1982); GP, Gunnison Plateau after Stanley and Collinson (1979); PA, Price area (Fig. 3 after Peterson 1976 and Chapman 1982); PR, P. R. Spring area (Picard and High 1970); RR, Raven Ridge area (Picard and High 1972); UF, Unita Formation (Middle to Upper Eocene) of central Uinta Basin (Bruhn et al. 1983).

attains a thickness of 500–600 m south of the Marysvale volcanic field (Fig. 2).

The paleodrainage leading from the San Luis uplift to the Uinta Basin was not the only instance of sediment dispersal over long distances within the Laramide region. Surdam and Stanley (1980) have documented the transport of volcaniclastic sediment southward from the Absaroka volcanic field in northwestern Wyoming as far as the Piceance Basin of northwestern Colorado (Fig. 2). Equivalents of the Uinta Formation within the Piceance Basin are largely volcaniclastic sandstones (Surdam and Stanley 1980). Feldspar-rich sandstones displaying westerly paleocurrent indicators in the Uinta Formation within the Uinta Basin (Fig. 10) may represent in part the most distal deposits of this northerly paleodrainage system, which could have entered the Uinta Basin by way of the Piceance Basin.

We conclude that systematic regional studies of Paleogene petrofacies and associated dispersal patterns within the Laramide region could prove to be a powerful tool for deciphering the timing of Laramide deformational episodes, and for understanding the evolving paleogeography of the Laramide region in real detail. In this sense, our study is only the beginning of a large task.

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