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BIOSTRATIGRAPHIC GUIDE TO UPPER DEVONIAN AND MISSISSIPPIAN ROCKS ALONG THE WASATCH FRONT AND CORDILLERAN HINGELINE, UTAH

By

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ABSTRACT

This paper provides a guide to measured sections of Upper Devonian (Famennian) and Mississippian (Kinderhookian, Osagean, and Meramecian) rocks visited on a three-day field trip between Provo, Utah, on the Wasatch Front and the Crawford Mountains in the Overthrust Belt along the Utah-Wyoming State line. The field trip, sponsored by the Pander Society, was held on April 30-May 2, 1978 in conjunction with the meeting of the Rocky Mountain section of the Geological Society of America at Provo, Utah. The rocks that may be sampled for conodonts are represented by the Middle to Upper *Scaphignathus velifer* (middle Famennian) through *Cavusgnathus* (middle Meramecian) conodont Zones. The measured sections and field trip provide a west to east traverse of the Sevier thrust system. The sections demonstrate the biostratigraphic, paleogeographic, and sedimentational differences among equivalent Famennian rocks as well as the transition from the Osagean and early Meramecian starved phosphatic basin on the west to time-equivalent carbonate rocks of the shelf margin on the east.

INTRODUCTION

This report describes the lithostratigraphy and biostratigraphy of nine important measured sections of Upper Devonian and Mississippian rocks that are located on an index map of northern Utah (fig. 1). These sections either were visited or were discussed (because of their proximity to visited sections) during the Pander Society-sponsored field trip held on April 30 - May 2, 1978, in conjunction with the meeting of the Rocky Mountain section of the Geological Society of America at Brigham Young University (BYU) in Provo, Utah. The described sections and the stop numbers assigned by the road log of the field trip (Sandberg, Petersen, and others, 1978) are:

(1) Rock Canyon (STOP 3) in the Wasatch Mountains behind the BYU campus.

(2) Flux (STOP 5) in the foothills at the northeast edge of the Stansbury Mountains.

(3) Broad Canyon (STOP 6) on Stansbury Island at the south end of Great Salt Lake.

(4) Porcupine Dam (STOP 7) in the southern part of the Bear River Range.

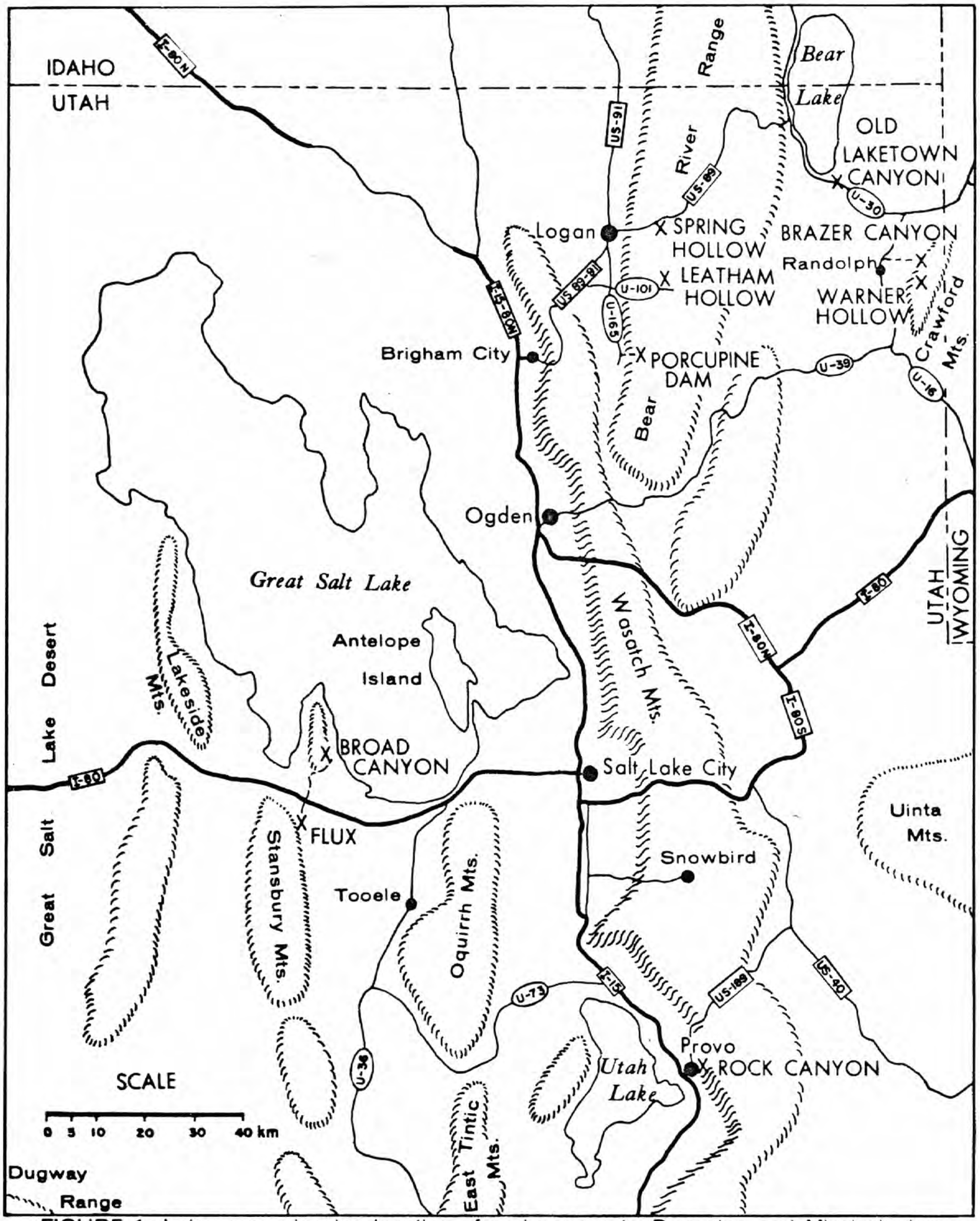


FIGURE 1.- Index map showing location of and access to Devonian and Mississippian stratigraphic sections.

(5) Leatham Hollow (not visited); type section of the Leatham Formation on Left Hand Fork of Blacksmith Fork Canyon in Bear River Range. (See Sandberg and Poole, 1977, fig. 11, for columnar section.)

(6) Spring Hollow (STOP 8) in Logan Canyon, Bear River Range.

(7) Old Laketown Canyon (STOP 9) at the south end of Bear Lake.

(8) Brazer Canyon (not visited), type locality of the Brazer Dolomite, on the west side of the Crawford Mountains. (See Sando and others, 1959, sec. 2 p. 2751-2754, for columnar section and description of rock units.)

(9) Warner Hollow (STOP 10), the easternmost locality for Mississippian phosphate, 5.5 km south of the mouth of Brazer Canyon in the Crawford Mountains.

Columnar sections for the seven localities that are not referenced in the above list are presented herein.

Conodont biostratigraphy and collecting localities are emphasized on the columnar sections because of the participation on the Pander Society field trip of conodont workers from outside the Great Basin area. However, brachiopods, corals, oncolites, goniatites, radiolarians, agglutinate foraminiferans, sponges, and trace fossils are also discussed and shown on the measured sections for the benefit of other participating biostratigraphers and geologists.

The paleogeographic and sedimentational treatment of Upper Devonian and Mississippian units places particular emphasis on two thin stratigraphic intervals, the conodont zonation of which is well documented. The large, diversified conodont faunas found in these intervals resulted from slow deposition. These faunas contrast sharply with the sparse conodont faunas of enclosing thick intervals of rapidly deposited quartzose and fossil-fragmental clastic rocks. The two thin intervals are shown by shading on the correlation charts (figs. 2 and 3). Because the vertical dimension of the correlation charts is based on conodont zonation (and hence, time) rather than on rock thicknesses, the thickness of these thin units, which contain several conodont zones, tends to be exaggerated. Consequently, the reader is cautioned to pay closer attention to the stated thicknesses of units rather than to their vertical dimensions.

The lower of the two emphasized intervals contains the Devonian-Mississippian boundary and includes most conodont zones between the Middle to Upper *Scaphignathus velifer* Zone and the Lower *Siphonodella crenulata* Zone (figs. 2 and 3). The complete Upper Devonian conodont zonation for the Western United

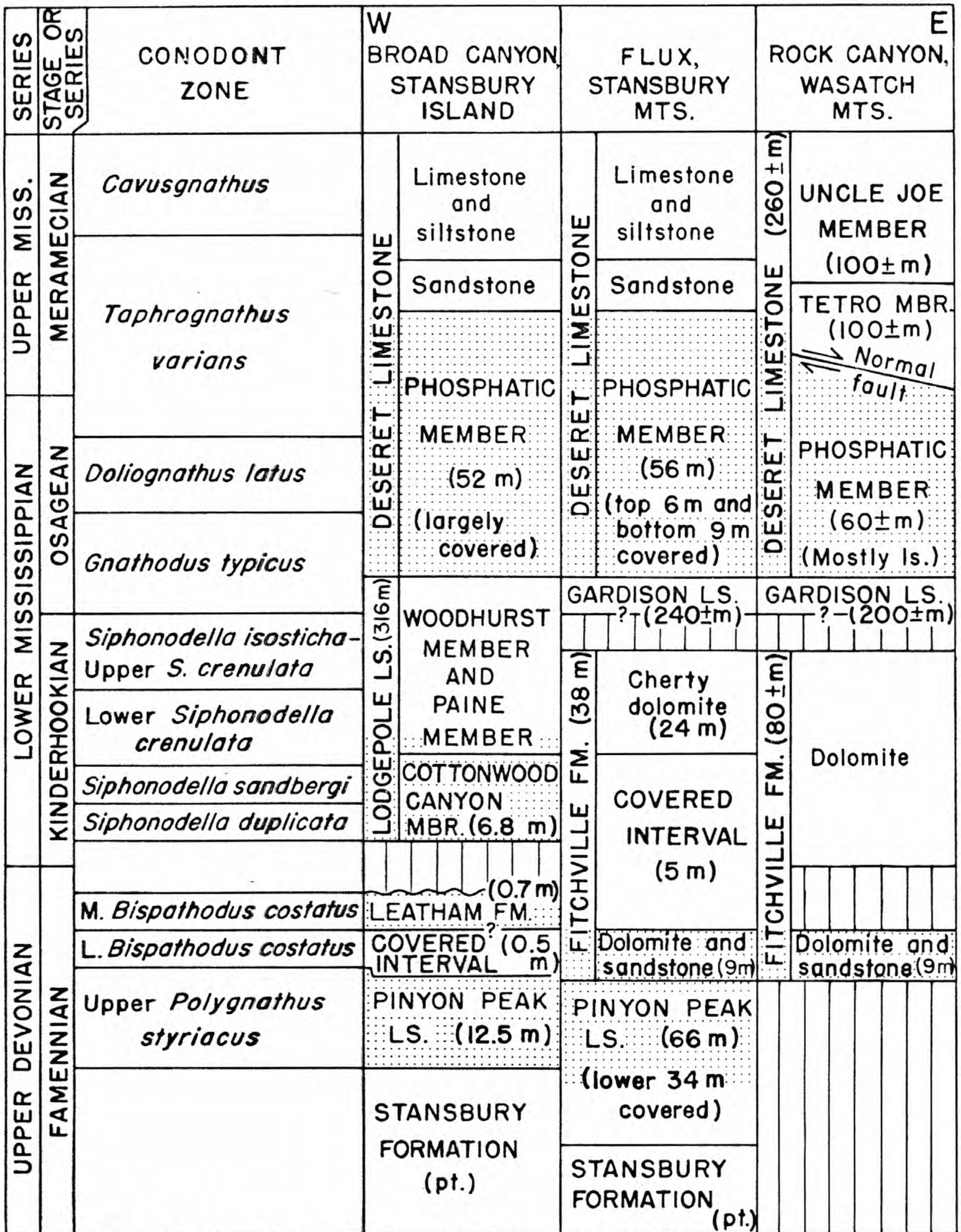


FIGURE 2.--Correlation chart for area south of Great Salt Lake. Stratigraphic intervals with well-documented conodont zonation are shaded.

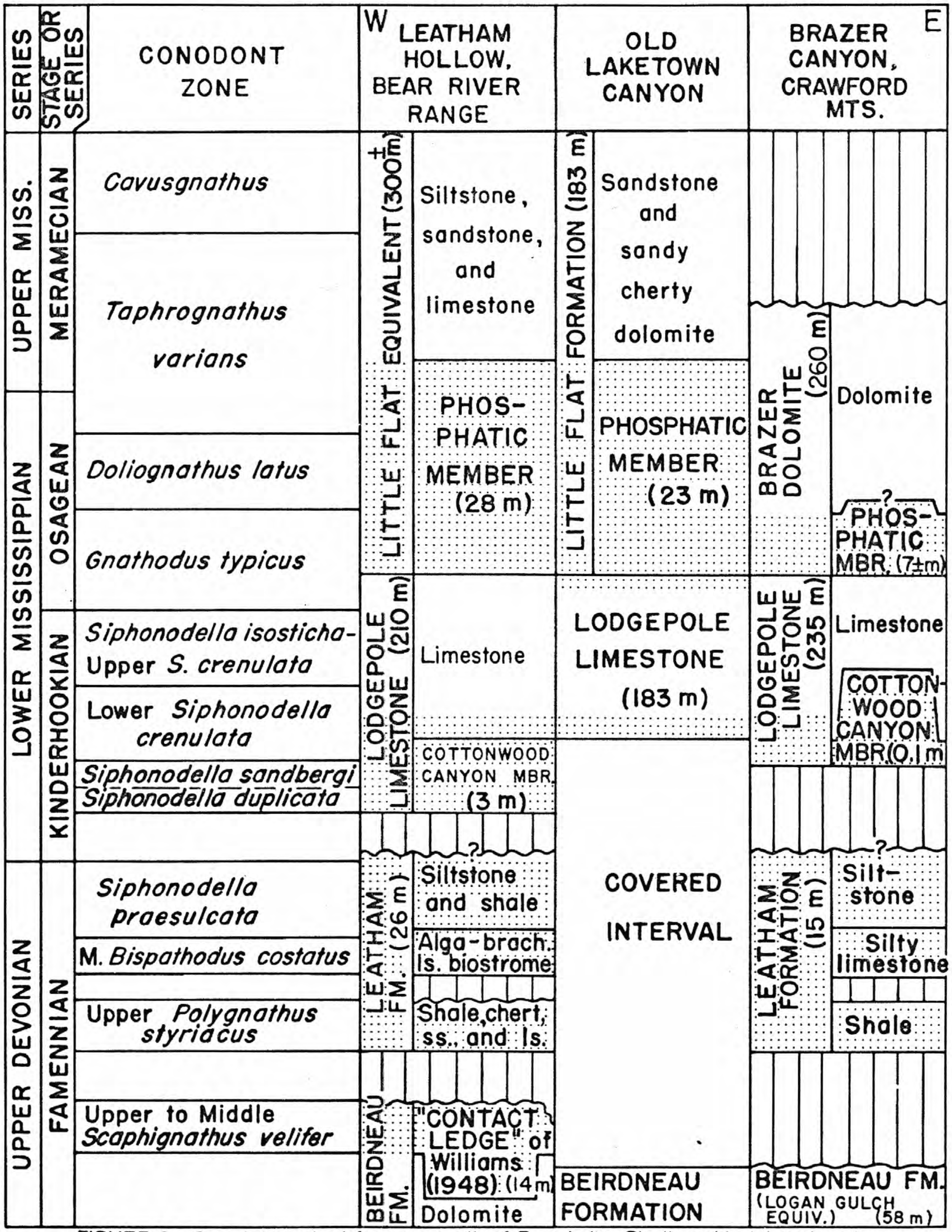


FIGURE 3.--Correlation chart for area south of Bear Lake. Stratigraphic intervals with well-documented conodont zonation are shaded.

States was shown by Sandberg and Poole (1977, fig. 2). The latest Devonian and Kinderhookian *Siphonodella* conodont zonation was proposed by Sandberg, Ziegler, and others (1978). Within the lower interval, the only zone that has not been found in the measured sections of this report is the *Siphonodella sulcata* Zone, which intervenes between the *S. praesulcata* and *S. duplicata* Zones (fig. 3). The *S. sulcata* Zone has been recognized, however, in the East Tintic Mountains (Sandberg and Poole, 1977, fig. 13), at the south edge of the report area (fig. 1). The lower interval comprises the Pinyon Peak Limestone (Upper Devonian) and Leatham Formation (Upper Devonian and Lower Mississippian?); the lower and middle parts of the Fitchville Formation (Devonian and Mississippian); and the basal, Cottonwood Canyon Member and lower beds of the overlying Paine Member and equivalents of the Lodgepole Limestone (Lower Mississippian).

The upper emphasized interval spans the *Gnathodus typicus* and *Doliognathus latus* Zones and the lower part of the *Taphrognathus varians* Zone (figs. 2 and 3; Sandberg, 1978). This stratigraphic interval constitutes a continuous phosphatic member of Mississippian (Osagean to early Meramecian) age, which is assigned at different localities to the base of various formations. In the measured sections of this report, the phosphatic member is assigned to the Deseret Limestone, Brazer Dolomite, Little Flat Formation, or an equivalent of the Little Flat Formation (figs. 2 and 3). In the Dugway Range and Lakeside Mountains (fig. 1) and farther west in Utah, the phosphatic member constitutes the basal part of the Woodman Formation (Lower and Upper Mississippian).

The thick carbonate interval, which intervenes between the Cottonwood Canyon Member of the Lodgepole Limestone and equivalents and the phosphatic member of the Deseret Limestone and equivalents, comprises the upper part of the Fitchville Formation, the Gardison Limestone, and the equivalent Paine and Woodhurst Members of the Lodgepole. Excluding the basal beds of the Paine Member and equivalents, which contain abundant faunas of the Lower *Siphonodella crenulata* Zone, this carbonate interval contains mainly sparse faunas of the Lower *S. crenulata*, *S. isosticha*-Upper *S. crenulata*, and *Gnathodus typicus* Zones. The reason for the sparsity of these faunas (yields of 0-20 conodonts/kg) is the rapid rate of sedimentation. This carbonate interval is herein calculated to have been deposited at a rate of 43-103 m/m.y. overall and at a rate of 200-275 m/m.y. for the upper part (Gardison Limestone and equivalents). In sharp contrast, the thin underlying and overlying intervals with highly abundant conodont faunas (yields as great as 200-1,000 conodonts/kg) were deposited much more slowly. A rate of 1-3 m/m.y. is calculated for the Cottonwood Canyon Member, and an overall rate of about 9-30 m/m.y. is calculated for the phosphatic member (8.9-11.5 m/m.y. for basinal shale and 26-30 m/m.y. for slope limestone).

The Stansbury Formation, which underlies the Pinyon Peak Limestone, and the Tetro and Uncle Joe Members and equivalents of the Deseret Limestone, which overlie its basal phosphatic member, were deposited rapidly. The Stansbury, which consists mainly of conglomerate and quartzite, has not been dated by conodonts. Its age is inferred to be Late Devonian (Frasnian and Famennian), however, because it unconformably overlies rocks as young as the Guilmette Formation (Middle and Upper Devonian) and is overlain with apparent conformity by the Pinyon Peak (Famennian). The Tetro and Uncle Joe Members and their equivalents, which are of Late Mississippian (Meramecian) age, contain sparse conodont faunas of the *Taphrognathus varians* and *Cavusgnathus* Zones.

STRATIGRAPHIC GLOSSARY

To make this biostratigraphic guide more understandable for the user who is unfamiliar with the Upper Devonian and Mississippian stratigraphic nomenclature of northern Utah, a brief summary of relevant stratigraphic units is presented in approximate chronologic order.

STANSBURY FORMATION.--This formation, named and described by Stokes and Arnold (1958), is limited in distribution to the Stansbury Mountains and to Stansbury Island, where the Broad Canyon section is located (fig. 1). It has a maximum thickness of about 520 m near Flux and consists mainly of conglomerate and quartzite with some interbedded silty sandy dolomite. The age is Late Devonian (Frasnian to Famennian). The Stansbury unconformably overlies Cambrian to Upper Devonian rocks and is overlain conformably by the Pinyon Peak Limestone.

BEIRDNEAU FORMATION.--Depositionally related to the Stansbury Formation and at least partly equivalent in age, the Beirdneau Formation, which was named by Williams (1948) as a member of the Jefferson Formation and later raised to formation status by Williams (1971), occurs in a large area of northern Utah. The southern boundary of its area of distribution is approximately a line drawn between the southern tips of the Lakeside and Crawford Mountains (fig. 1). The formation is as much as 330 m thick and consists mainly of restricted-marine silty and sandy calcitic dolomite with salt casts and some interbedded dolomitic sandstone. It was named for Beirdneau Peak on the north side of Logan Canyon, northeast of Spring Hollow (fig. 1). An accessible reference section was described by Williams (1971) on Blacksmith Fork, 6 km southwest of Leatham Hollow (fig. 1). The Beirdneau overlies the Middle and Upper Devonian Hyrum Dolomite with apparent conformity and is unconformably overlain by the Leatham Formation.

The bulk of the Beirdneau Formation is laterally continuous with the Logan Gulch Member of the Three Forks Formation of southwestern Montana (Sandberg and Poole, 1977, p. 159 and fig. 4B, lithofacies 2 and 3). The upper 10 m, however, is a lithologically distinct cliff-forming limestone (fig. 11) that was called the "contact ledge" by Williams (1948). This thin unit, which grades upward from restricted-marine pelmicrite to open-marine nodular argillaceous limestone, correlates with the Trident Member of the Three Forks Formation in Idaho and southwestern Montana (Sandberg and Poole, 1977, p. 164-166). The upper part of the "contact ledge" yields conodont faunas containing a new morphotype of *Palmatolepis distorta*; these faunas represent the polygnathid-icriodid and polygnathid-pelekysgnathid biofacies of the Middle to Upper *Scaphignathus velifer* Zone.

<u>PINYON PEAK LIMESTONE</u>.--This formation was named in the Tintic mining district in 1919. The type locality, Pinyon Peak, is in the East Tintic Mountains at the south edge of the index map (fig. 1). The Pinyon Peak occupies the area south of a line extending from the south tip of the Lakeside Mountains to the northeast corner of the Uinta Mountains (fig. 1). The Mowitza Shale, as used by Sandberg (1976, fig. 1), is abandoned (Sandberg and Poole, 1977) and is now included in the Pinyon Peak. The regional distribution of the Pinyon Peak and a stratigraphic section at its type locality were shown by Sandberg and Poole (1977, figs. 10, 12, 13). The Pinyon Peak, which has a maximum thickness of about 115 m, comprises two members. The lower member, which consists mainly of restricted-marine pelmicrite and intraformational limestone breccia, is largely covered at Flux (fig. 7) and is absent at Broad Canyon (fig. 9). The upper member, which crops out at both localities, consists mainly of open-marine fossiliferous nodular limestone at Flux and fossiliferous silty mudstone at Broad Canyon; elsewhere these lithologies are interbedded (Sandberg and Poole, 1977, fig. 14). In the report area, the Pinyon Peak conformably overlies the Stansbury Formation and is conformably overlain by the Fitchville Formation.

The lower member of the Pinyon Peak Limestone, which is tentatively correlated with the "contact ledge" of the Beirdneau Formation mainly contains sparse conodont faunas of the polygnathid-icriodid biofacies and, near the top, a new *Pedavis*-like species with two lateral processes that is tentatively assigned to *Icriodus*. The upper member correlates with the lower part (units 1-3 of Sandberg and Gutschick, 1969) of the Leatham Formation. It contains abundant faunas of the polygnathid-icriodid biofacies of the Upper *Polygnathus styriacus* Zone, as well as horn corals and the brachiopods *Paurorhyncha, Gastrodetoechia,* and *Cyrtospirifer*. The horn corals are comparable to, but slightly older than, those found in the Etroeungt Limestone of Belgium.

LEATHAM FORMATION.--The Leatham Formation was named by Holland (1952) for its type locality at Leatham Hollow (fig. 1). Its distribution is restricted mainly to the Bear River Range and Crawford Mountains (fig. 1). Its thickness is 26 m at the type section, which is described by Sandberg and Poole (1977, fig. 11), and 20.5 m nearby at Spring Hollow (fig. 11). The type Leatham was divided into seven units by Sandberg and Gutschick (1969); all except unit 2 are present at Spring Hollow. The lower three units, 1-3, consist mainly of radiolarian- and spicule-bearing black siltstone, mudstone, and chert with basal beds of conodont-rich lag sandstone and limestone. Unit 4, which overlies an unconformity, consists in ascending order, of a lag sandstone, a conchostracan- and *Tasmanites*-rich shale, and a ledge-forming dark pyritic fossiliferous limestone. Unit 5 is silty limestone with abundant oncolites and brachiopods and fewer sponges; it represents a segment of a widespread but narrow alga-brachiopod biostrome that extends from Montana to Nevada. Unit 6 is silty limestone and calcareous siltstone containing agglutinate foraminiferans, trace fossils, and brachiopods. Unit 7 is black nonfossiliferous carbonaceous mudstone. The Leatham Formation overlies a major regional unconformity, which in the report area is mainly above the Beirdneau Formation. The Leatham is overlain unconformably by the Cottonwood Canyon Member of the Lodgepole Limestone.

The age of the Leatham Formation is Late Devonian and Early Mississippian(?). It was once depositionally continuous with the Sappington Member of the Three Forks Formation in western Montana and east-central Idaho and with the middle member of the Pilot Shale in western Utah and eastern Nevada (Sandberg, 1976; Sandberg and Poole, 1977). The thin, dark limestone that overlies the Pinyon Peak Limestone at Broad Canyon (fig. 9) is correlated with unit 4 of the Leatham. Units 1-3 of the Leatham, which contain the palmatolepid-polygnathid biofacies of the Upper *Polygnathus styriacus* Zone, are the western, deeper water time-equivalent of the shallower water Pinyon Peak Limestone and basal beds of the Fitchville Formation. Units 4-7 of the Leatham, however, represent the time missing at an unconformity between the lower and middle parts of the Fitchville. Throughout the Bear River Range, unit 4 of the Leatham yields an abundant, diverse fauna, listed by Sandberg, Streel, and Scott (1972), of the palmatolepid-polygnathid biofacies of the Middle *Bispathodus costatus* Zone. Unit 5 yields faunas of the overlying *Siphonodella praesulcata* Zone, as well as abundant brachiopods, including *Syringothyris, Spirifer,* and *Rhipidomella.* At Porcupine Dam (fig. 10),both the Middle *B. costatus* and *S. praesulcata* Zones are present in a ledge, only 0.7 m thick, that combines units 4 and 5 (Sandberg and others, 1972, p. 186). Unit 6 contains sparse, undiversified faunas of the *S. praesulcata* Zone. Unit 7, which has not yet yielded conodont faunas, might be Mississippian.

FITCHVILLE FORMATION.--This formation was named by Morris and Lovering (1961) for exposures on Fitchville Ridge in the East Tintic Mountains (fig. 1). In their written stratigraphic section, they identified 10 informal lithologic units within the Fitchville Formation. Some of these have proven to be widely distributed marker beds, which are useful in mapping the regional extent of the formation. The area of distribution of the Fitchville is south of a line from the north tip of the Stansbury Mountains to the central part of the Wasatch Mountains, east of Salt Lake City (fig. 1). Within this area, the distribution of the distinctive "curly limestone," which normally caps the formation, was shown by Proctor and Clark (1956). A bed of sandstone or grit marks the base of the Fitchville at most localities. A stratigraphic section by Sandberg and Poole (1977, fig. 13) showed the conodont zonation based on 22 samples of the lower and middle parts of the type Fitchville. The Fitchville has a maximum thickness of about 88 m in the type locality. The Devonian lower part of the formation, 34 m thick, is mainly fossil-fragmental limestone and encrinite with common horn corals. The middle part, 14 m thick, is mainly thin-bedded argillaceous limestone that lithologically resembles the Kinderhookian Paine Member of the Lodgepole Limestone. The upper part, 40 m thick, is largely massive dolomite with common corals, capped by thin units of shallow-marine birdseye and algal (curly) limestone. Elsewhere, as in the Wasatch Mountains, the Fitchville is almost entirely dolomitized and is largely featureless. The Fitchville conformably overlies the Pinyon Peak Limestone and is disconformably overlain by the Gardison Limestone, which visibly channels across the two highest units of the Fitchville in some outcrops.

The age of the Fitchville Formation is Late Devonian (late Famennian) and Early Mississippian (Kinderhookian). The late Famennian lower part mainly contains faunas of the Lower *Bispathodus costatus* Zone, but the lowest 2-3 m contains the same fauna of the polygnathid-icriodid biofacies of the Upper *Polygnathus styriacus* Zone that characterizes the underlying upper member of the Pinyon Peak Limestone. The coral faunas of the lower part of the Fitchville and upper member of the Pinyon Peak are being studied by W. J. Sando. They include advanced clisiophyllid and caninoid corals comparable to coral faunas in the Etroeungt Limestone in Belgium. Similar forms also occur in the Viséan (Meramecian equivalent) in Belgium, but they generally do not occur in Kinderhookian, Osagean, and Meramecian rocks in North America. The early Kinderhookian middle part of the Fitchville is separated from the lower part by an unconformity, evidenced faunally by the absence of the Middle and Upper *B. costatus* and *Siphonodella praesulcata* Zones. The middle part contains mainly the *S. sulcata* to *S. sandbergi* Zones, but the *S. duplicata* Zone has not been documented faunally. The Lower *S. crenulata* Zone occurs at about the boundary between the middle and upper parts, and in different areas may occur in either or both parts. The faunas of the late Kinderhookian upper part have not been extensively studied; they generally are poorly preserved because of pervasive dolomitization.

LODGEPOLE LIMESTONE.--This formation has its type locality in Montana, where it is included in the Madison Group. The type sections of the Lodgepole and Madison Limestones have been described by Sando and Dutro (1974). The Lodgepole was extended into the Crawford Mountains, Utah, by Sando, Dutro, and Gere (1959) and subsequently has been widely used in the Bear River Range (e.g., Mullens and Izett, 1964). The type Lodgepole consists, in ascending order, of the Kinderhookian Cottonwood Canyon and Paine Members and the Osagean Woodhurst Member. The Paine and Woodhurst Members have not been recognized in the Bear River Range, where the lower and upper cliff-forming parts of the Lodgepole are known informally as the "Chinese Wall" and "Upper Chinese Wall" (Holland, 1952, p. 1723). Herein, with the assistance of W. J. Sando, the Paine and Woodhurst are recognized at Broad Canyon (fig. 1). The thin, basal Cottonwood Canyon Member was named in Wyoming by Sandberg and Klapper (1967), who discussed its distribution, complex stratigraphic relations, conodont faunas, and conodont zonation. In Utah, the Cottonwood Canyon was first recognized by Sandberg and Gutschick (1969), and the member is almost invariably present wherever the Lodgepole is identified. The Lodgepole ranges in thickness from 210 to 316 m, and the Cottonwood Canyon ranges from 0.1 to about 7 m. The Cottonwood Canyon consists mainly of sandstone and limestone lag beds that are extremely rich in conodonts, interbedded, where the member is thickest, with pale-red-purple mudstone and siltstone and (or) black carbonaceous shale. The Paine, except for the basal 5 m, which is conodont-rich encrinite and fossil-fragmental limestone, consists of thin-bedded, argillaceous, cherty slope limestones containing sparse conodont faunas. The Woodhurst, which comprises beds of the same lithology interbedded with encrinitic and bioclastic limestones, contains abundant coral faunas at Broad Canyon. The Cottonwood Canyon Member rests unconformably on the Leatham Formation and truncates it southward within the Bear River Range. The Lodgepole is overlain conformably by the phosphatic members of the Deseret Limestone, Little Flat Formation and equivalents, and Brazer Dolomite.

The age of the Lodgepole Limestone is Kinderhookian and early Osagean. The Cottonwood Canyon Member in Utah contains the *Siphonodella duplicata, S. sandbergi,* and Lower *S. crenulata* Zones. The basal 5 m of the Paine Member or of equivalent undifferentiated limestones contains abundant faunas of the Lower *S. crenulata* Zone. The middle part of the Lodgepole contains sparse faunas of the same zone and of the *S. isosticha*-Upper *S. crenulata* Zone, but the zonal boundary has not been located. The upper part of the Lodgepole contains faunas, without *Siphonodella*, of the early Osagean *Gnathodus typicus* Zone.

GARDISON LIMESTONE.--The Gardison Limestone was named by Morris and Lovering (1961) for a spur in the northern part of the East Tintic Mountains (fig. 1). The Gardison has virtually the same distribution as the Fitchville Formation in the area south of a line between the north tip of the Stansbury Mountains and the middle part of the Wasatch Mountains, east of Salt Lake City (fig. 1). The type Gardison, as described by Morris and Lovering (1961, p. 90), is about 150 m thick. The formation ranges from 200 to 240 m in thickness in areas to the north. The Gardison consists mainly of shallow-water, medium- to dark-gray, cherty and noncherty, coarsely fossil-fragmental limestone and encrinite, which contain abundant, well-preserved coral faunas, assigned by W. J. Sando to his C_1 coral zone, as well as gastropods and brachiopods. The basal bed commonly is crossbedded encrinite, which channels into the underlying Fitchville Formation. The bulk of the Gardison represents rapidly accumulated bioclastic deposits of the carbonate platform or shelf margin. However, the highest bed, less than 5 m thick, is generally a very dark gray deep-water micritic limestone with a different megafauna. This highest bed of the Gardison is transitional with and genetically related to the overlying phosphatic member of the Deseret Limestone.

The Gardison Limestone is considered to be entirely Early Mississippian (early Osagean) in age. The C_1 coral zone, elsewhere ranges from Kinderhookian into Osagean, but the Kinderhookian index conodont, *Siphonodella*, has not been found in the Gardison. Nondiagnostic shallow-water conodont faunas are extremely sparse because of the rapid rate of sedimentation. The highest bed, however, contains the *Gnathodus typicus* Zone, as do overlying basal beds of the phosphatic member. The Gardison correlates with the Woodhurst Member of the Lodgepole Limestone.

<u>DESERET LIMESTONE</u>.--This formation, named by Gilluly (1932, p. 22-25), has its type locality on the west side of the Oquirrh Mountains, south of Tooele (fig. 1). It comprises three members, in ascending order: a phosphatic member; the Tetro Member, which was introduced in early mining reports; and the Uncle Joe Member, which was named and described by Morris and Lovering (1961, p. 95-96) in the East Tintic Mountains (fig. 1). The Deseret Limestone is recognized at Broad Canyon as well as in the mountain ranges south of a line from the north tip of the Stansbury Mountains to the central part of the Wasatch Mountains, east of Salt Lake City (fig. 1). In the report area, it has a maximum thickness of about 335 m in the East Tintic Mountains. The phosphatic member, as much as 60 m thick, comprises mainly dark-colored beds of organic phosphatic shale, siltstone, and mudstone (enclosing large calcareous concretions or bullions); pelletal, peloidal, and oolitic phosphorite; chert; partly cherty micritic limestone; and some light-colored interbeds of debris-flow limestone. Both types of limestone predominate in eastern sections, such as Rock Canyon, where intervening shale beds are thin. The phosphatic member in the East Tintic Mountains has been described in some detail by Morris and Lovering (1961, p. 99-104). Several stratigraphic sections were illustrated by Gutschick (1976), and the age, correlation, distribution, and origin of the phosphatic member were discussed by Sandberg and Gutschick (1977a).

The other members of the Deseret Limestone have not been studied as intensively as the phosphatic member. The Tetro Member is a moderately deep- to shallow-water, lithologically highly variable, slope-deposited unit comprising siltstone, sandstone, and sandy cherty limestone. The Uncle Joe Member, which is a prograding carbonate buildup deposited on a gentle ramp or slope, consists mainly of fossil-fragmental limestone and encrinite and some micritic limestone. The coarser limestone beds generally become sandier and sandstone interbeds become more numerous and thicker upward, due to an increase in craton-derived clastics. The Deseret Limestone conformably overlies the Gardison or Lodgepole Limestone and is conformably overlain by the Upper Mississippian (Meramecian) Humbug Formation.

The age of the Deseret Limestone is Early and Late Mississippian (early Osagean to middle Meramecian). The phosphatic member is early Osagean to early Meramecian, as discussed by Sandberg and Gutschick (1977a), and contains a succession of faunas of the *Gnathodus typicus, Doliognathus latus*, and *Taphrognathus varians* Zones. The ammonoids *Beyrichoceras* and *Djhaprakoceras* occur at Flux (Petersen, 1969) and at Old Laketown Canyon (Sando and others, 1976) in concretions and limestones of the late Osagean part of the phosphatic member. The concretions also contain the brachiopod *Leiorhynchoidea carbonifera* as well as a microfauna comprising many types of radiolarians, agglutinate foraminiferans, conodonts, and sponge spicules (Gutschick and Sandberg, 1977, 1978). The micritic limestones of the member commonly contain small horn corals, and the siltstones contain *L. carbonifera*. The Tetro and Uncle Joe Members span parts of the early and middle Meramecian. The lowest occurrence of *Cavusgnathus* is believed to be within the upper part of the Deseret, which contains Mamet zone 13 and 14 foraminiferans and zone E corals (W. J. Sando, written commun., June 1977). The upper two members of the Deseret have not been systematically searched for conodonts, however, because of lithologies unfavorable for conodonts and because the corals, foraminiferans, and algae provide a more precise basis for dating and zonation.

BRAZER DOLOMITE.--This formation, which was named by Richardson (1913) for Brazer Canyon (fig. 1), was restudied and thoroughly described by Sando, Dutro, and Gere (1959). The Brazer Dolomite is restricted to the Crawford Mountains, although largely undolomitized rocks of the same age are present in the western Uinta Mountains (fig. 1). Most measured sections in the type area are complicated by thrust faults, but Sando, Dutro, and Gere (1959) determined that the Brazer has a consistent thickness of about 260 m. They recognized three members, numbered 1 to 3 in ascending order. Member 1, about 42 m thick, is dark-yellowish-brown-weathering fine-grained dolomite with black tabular and nodular chert composing about half the member. Member 2, about 86 m thick, is yellowish-gray-weathering medium- to coarsely-crystalline dolomite with light-colored chert composing about 30 percent of the member. Member 3, about 128 m thick, comprises interbeds of several types of dolomite, some of which are oolitic or sandy; it contains less chert than either of the lower two members. The Brazer Dolomite is largely a dolomitized bioclastic shelf limestone; the lower member, however, may represent deposition on a gentle slope or in somewhat deeper water. The Brazer rests conformably on the Lodgepole Limestone. The lower 5 m of the type Brazer is covered, but a thin bed of phosphorite is present at the base of this interval, where it is well exposed at Warner Hollow (fig. 13), about 5.5 km south of the mouth of Brazer Canyon. The Brazer is overlain unconformably by sandstone and carbonate rocks of Pennsylvanian age.

The Brazer Dolomite is Early and Late Mississippian (early Osagean to early Meramecian) in age. The megafaunas, comprising mainly poorly preserved corals, brachiopods, and bryozoans, were listed by Sando, Dutro and Gere (1959, p. 2755-2761). Sample REX-8 from a bed of dolomite just above the basal phosphorite at Warner Hollow (fig. 13) yielded a large, diversified conodont fauna of the *Gnathodus typicus* Zone. The Brazer Dolomite is the dolomitized equivalent of the Mission Canyon Limestone of the Madison Group in Montana.

LITTLE FLAT FORMATION.--This formation was named for Little Flat Canyon in the Chesterfield Range of southeastern Idaho by Dutro and Sando (1963). They correlated the Little Flat Formation southward to Old Laketown Canyon (fig. 1) and applied that name to beds previously called members A and B of an unnamed formation by Sando, Dutro, and Gere (1959, fig 5). Sando, Dutro, Sandberg, and Mamet (1976, fig. 2) revised the age of the Little Flat and reexamined the correlation between Little Flat and Old Laketown Canyons; they recognized a basal phosphatic silty member at both localities. Although the Little Flat has not been formally recognized elsewhere in Utah, the lower beds of a thick sequence that has been erroneously called "Brazer Limestone" in the Bear River Range are herein recognized to be an equivalent of the Little Flat. At Old Laketown Canyon the Little Flat is 244 m thick. It comprises a phosphatic member, 23 m thick, consisting of phosphatic mudstone and siltstone with thick interbeds of micritic limestone and dolomite and thin interbeds of phosphorite; and an upper unit, 221 m thick, of sandy cherty dolomite and sandy dolomitic siltstone. The Little Flat and equivalents conformably overlie the Lodgepole Limestone and are conformably overlain by the Meramecian Monroe Canyon Limestone and equivalents.

The Little Flat Formation is Early and Late Mississippian (early Osagean to middle Meramecian) in age. It contains at the base the *Gnathodus typicus* conodont zone and at the top zone E corals and Mamet zone 13-14 foraminiferans. Hence the Little Flat is an approximate time equivalent of the Deseret Limestone. Moreover, the phosphatic member is almost identical in age and lithology to and laterally continuous with the phosphatic member of the Deseret. This member is also equivalent in age to the entire Brazer Dolomite, which is 11 times as thick (260 m vs. 23 m) just 22 km to the southeast. At Old Laketown Canyon, the carbonate

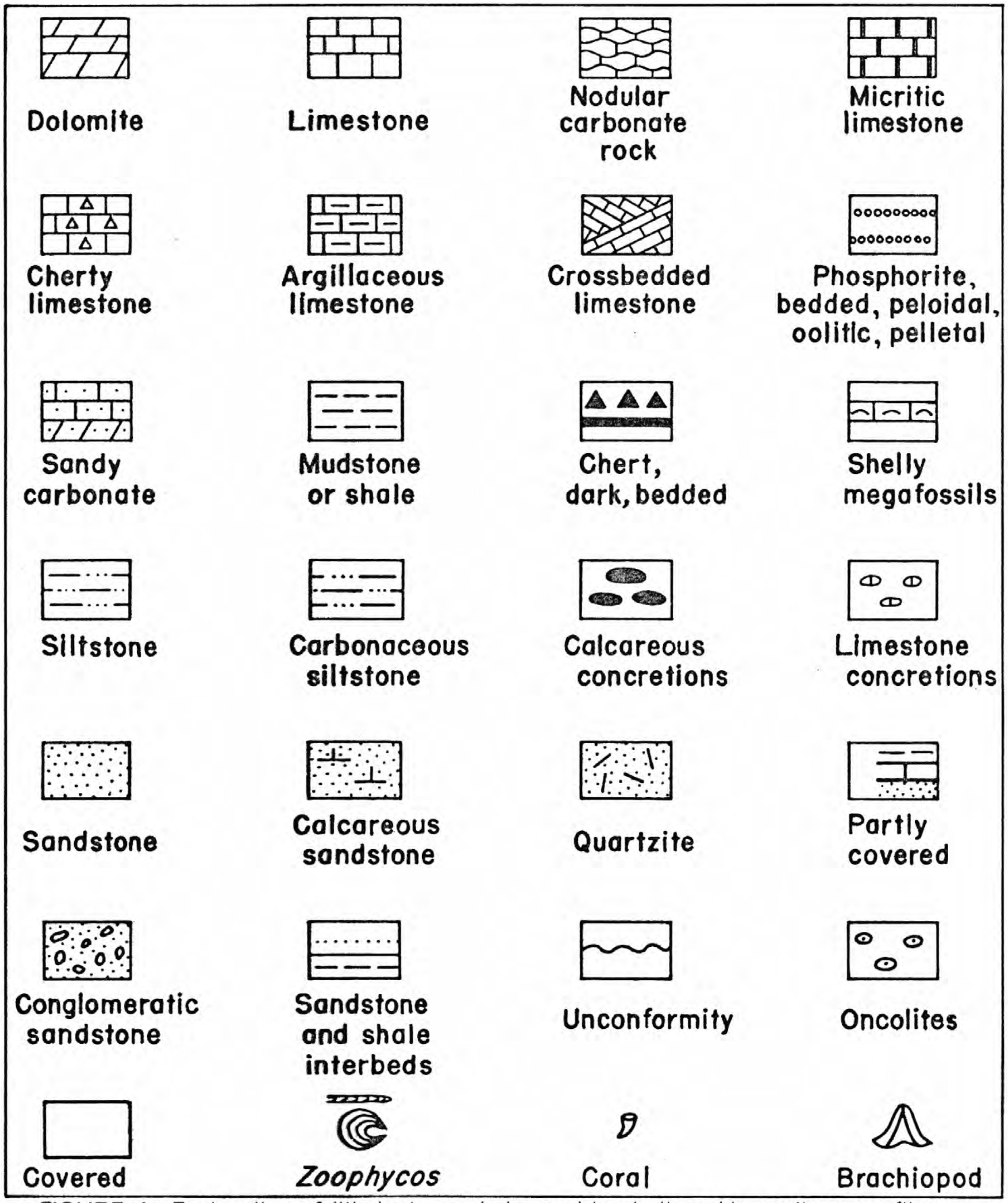


FIGURE 4.--Explanation of lithologic symbols used in stratigraphic sections, profiles, and models.

interbeds of the phosphatic member have yielded mainly sparse faunas of the *Doliognathus latus* Zone, but the lowest shale beds are undated.

STRATIGRAPHIC SECTIONS

The stratigraphic sections are discussed in the order in which they are visited or passed by the Pander Society field trip rather than in progression from basin to shelf. The lithologic symbols used on these sections are explained by figure 4. The sections give Sandberg's conodont sample numbers and his determinations of conodont color-alteration-index (CAI) values, according to the scheme presented by Epstein, Epstein, and Harris (1977). The CAI values indicate the range of maximum temperatures to which the host rocks have been subjected.

ROCK CANYON (STOP 3).--The Devonian to Mississippian sequence that is viewed while walking up the trail eastward from the mouth of Rock Canyon is mapped on the Bridal Veil Falls geologic quadrangle map (Baker, 1972). An excellent exposure of the contact between the Middle Cambrian Maxfield Limestone and the Devonian lower part of the Fitchville Formation can be seen about 25 m above the trail on the north wall. The basal clastic unit of the Fitchville forms a 3.2-m-thick yellowish-gray-weathering ledge that is largely quartz grit, conglomerate, sandstone, and siltstone with a little silty dolomite at the base and top. The dolomite at the base yielded a soon-to-be-named new species of *Icriodus* (which was called *I.* cf. *I. costatus* by Sandberg, 1976) and *Polygnathus semicostatus*. This fauna probably is referable to the polygnathid-icriodid biofacies of the Upper *Polygnathus styriacus* Zone. The overlying cliff-forming unit is medium-light-gray dolomite, 6 m thick, that yielded conodont faunas with *P. extralobatus*. This species occurs in the highest part of the Upper *P. styriacus* Zone and in the overlying Lower *Bispathodus costatus* Zone. The unit may be sampled just above creek level. All conodonts from this part of the Fitchville have a CAI value of 6.

Beyond this exposure, the canyon narrows in the upper part of the Fitchville Formation and Gardison Limestone and then widens abruptly in the lower part of the Deseret Limestone. Looking high onto the north wall from the trail, one can see the profile of the Deseret shown on figure 5. This is the area where the stratigraphic section (fig. 6) was measured. All conodonts from this part of the sequence have CAI values of 5, which indicate temperatures of >300°C (Epstein and others, 1977).

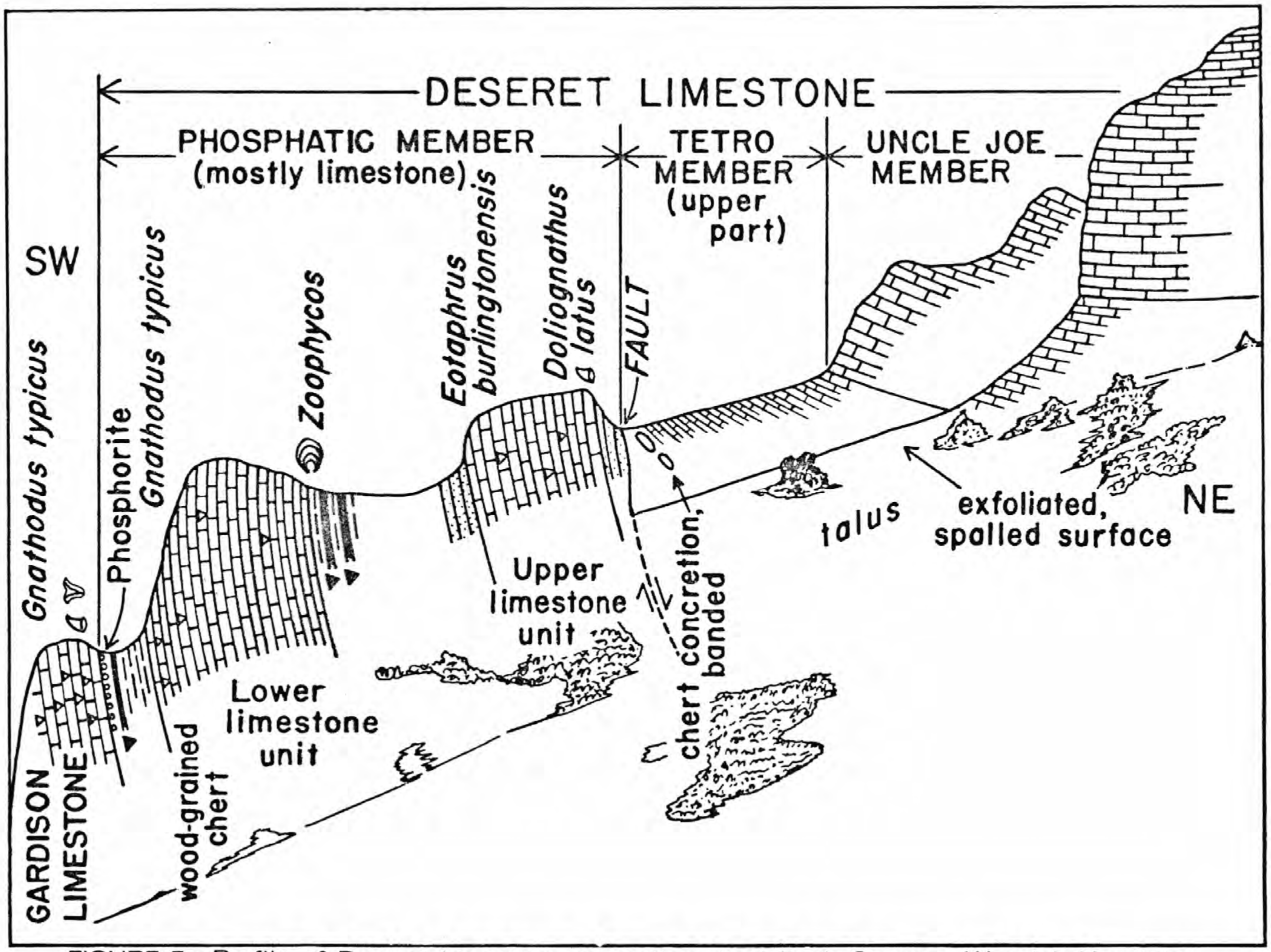


FIGURE 5.--Profile of Deseret Limestone on north wall of Rock Canyon, Wasatch Mountains. (STOP 3)

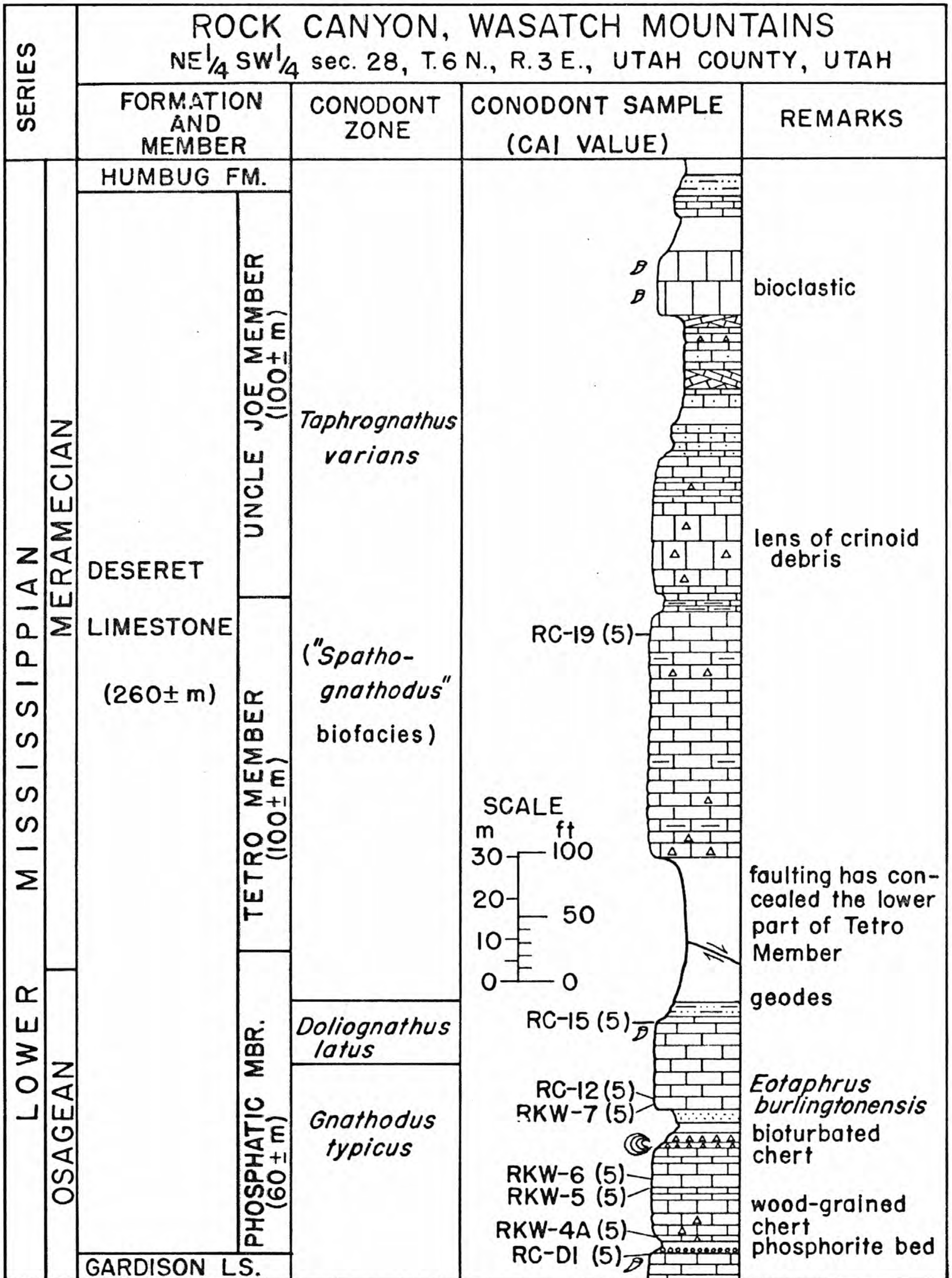


FIGURE 6.--Stratigraphic section of Deseret Limestone and adjacent units, Rock Canyon, Wasatch Mountains. (STOP 3)

The thin beds of shale, mudstone, and siltstone that compose the base, middle, and top of the phosphatic member of the Deseret Limestone, as seen in the profile (fig. 5), are covered by limestone talus on the lower slopes. However, the thick lower and upper limestone units extend far downslope and may be examined not too high above creek level. The lower unit is uniformly dark-gray micrite with the thin, even, planar bedding that characterizes slope limestones of the phosphatic member. This unit yielded only sparse faunas, dominated by *Polygnathus communis communis*, of the *Gnathodus typicus* Zone. The upper unit contains similar slope limestones with common interbeds of fine to coarse encrinitic debris-flow limestone that channel into the slope limestones. The lower part of this unit yielded conodont faunas with *Eotaphrus burlingtonensis*, and the upper part yielded faunas with *Doliognathus latus*. *Eotaphrus burlingtonensis* has not been found below *D. latus* in Belgium nor in the Upper Mississippi Valley. Bioturbated chert with *Zoophycos* burrows may be found in the talus between the two limestones. Higher on the slope, this chert occurs in place on top of the lower limestone unit. A normal fault obscures the stratigraphic relations between the phosphatic member and overlying Tetro Member.

FLUX (STOP 5).--This section in the low foothills on the northeast side of the Stansbury Mountains, near Flux siding, presents an unparalleled opportunity to walk across excellent, low exposures of the sequence from the Stansbury Formation to the Humbug Formation. Only four major intervals of cover that result from alluvial fill on benches of the Provo level of glacial Lake Bonneville mar the otherwise continuous exposures. One of these, unfortunately, is on the lower part of the phosphatic member of the Deseret Limestone, and this prevents proposing Flux as the type locality for a formal member. The geology of this area was shown on a planimetric map of the Timpie 15-minute quadrangle by Rigby (1958, pl. 1). Examination of the sequence begins with a conglomeratic unit in the upper part of the Stansbury Formation, as shown on figure 7, and continues upsection.

The conglomerate of the Stansbury Formation in the outcrop north of the dirt road contains cobbles and boulders of carbonate rocks derived mainly from Silurian and Devonian formations. This is overlain by sandstones, which cap the Stansbury and dip eastward about 55° into a largely covered interval in the lower part of the Pinyon Peak Limestone. Just over the crest of the saddle to the north is an isolated exposure of a thin bed of limestone in the Pinyon Peak that contains in association with *Palmatolepis rugosa* and *P. perlobata postera* a new species of

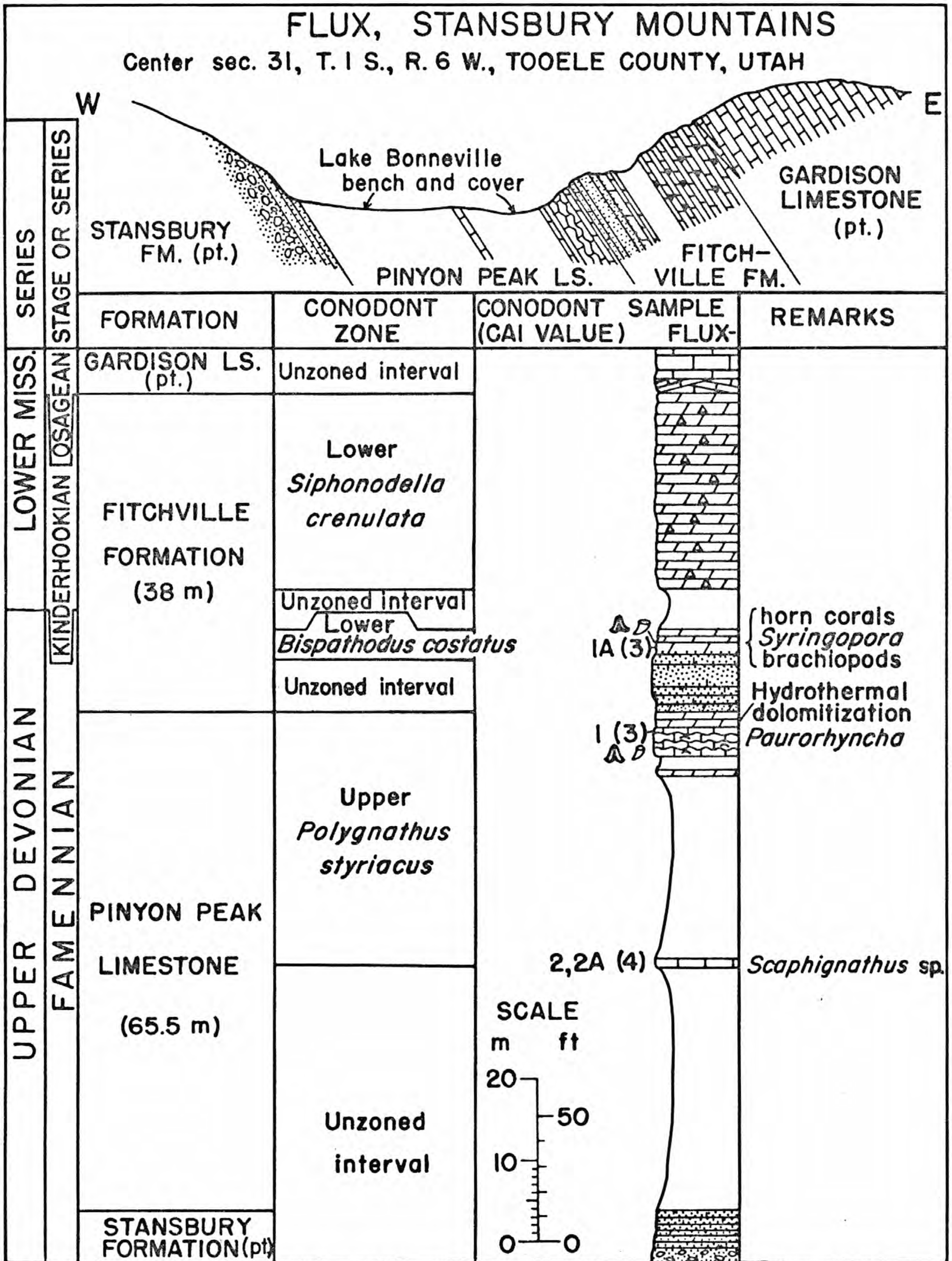


FIGURE 7.--Stratigraphic section and profile of Pinyon Peak Limestone, Fitchville Formation, and adjacent units near Flux, Stansbury Mountains. (STOP 5)

Scaphignathus (Sandberg, 1976) that resembles *Cavusgnathus* except for having an inverted basal cavity and a small pit. The next higher exposures are of the upper part of the Pinyon Peak, which consists of mottled "leopard spot" nodular fossil-fragmental limestone containing the brachiopod *Paurorhyncha* and a conodont fauna of the polygnathid-icriodid biofacies of the Upper *Polygnathus styriacus* Zone that is dominated by *Polygnathus* and contains less than 1 percent of *Palmatolepis* stragglers (Sandberg, 1976). Hydrothermal dolomitization cuts irregularly across the limestone, which is entirely altered to dolomite a few meters higher.

Resting on the Pinyon Peak Limestone is the quartzitic sandstone and grit marker bed at the base of the Fitchville Formation. There is slight evidence of burrow and crevice filling at the base of this bed, but the contact between it and the highest bed of the Pinyon Peak, which contains scattered quartz grit and pebbles, is essentially gradational. The sandstone bed grades upward into sandy bioclastic dolomite that contains a fauna of horn corals, comparable to those in the Etroeungt Limestone in Belgium. This dolomite contains a conodont fauna of the Lower *Bispathodus costatus* Zone that includes *B. costatus, B. aculeatus, Icriodus costatus, Polygnathus extralobatus,* and *P. delicatulus*. This fauna is similar to faunas from the lower part of the Madison Limestone in Wyoming (Sandberg and Klapper, 1967, p. B51, B57). A short covered interval separates this dolomite from a unit of *Siphonodella*-bearing cherty dolomite that caps the formation. The cherty dolomite is similar in stratigraphic position and lithology to cherty limestone in the basal part of the Paine Member of the Lodgepole Limestone just to the north on Stansbury Island. However, the cherty beds here are very thin, as is the entire Fitchville. Thinning is attributed at least in part to truncation by the overlying Gardison, which differs lithologically from beds of the Lodgepole in the same position on Stansbury Island.

As at most other localities, the basal bed of the Gardison Limestone is a crossbedded calcarenite or encrinite with abundant corals. The overlying beds of the Gardison constitute a monotonous sequence of coral-bearing limestone and dolomite, which we estimate to be about 240 m thick and predict to contain only sparse conodont faunas. The highest beds of the Gardison Limestone form a steep dip-slope cliff separated from the thick ledge of dark-gray micritic limestone in the phosphatic member of the Deseret Limestone by a covered stratigraphic interval about 10 m thick, as shown in the profile of this member (fig. 8).

To continue the examination of this section, one must descend into the gulch to see some of the fresh exposures that were made during exploration for

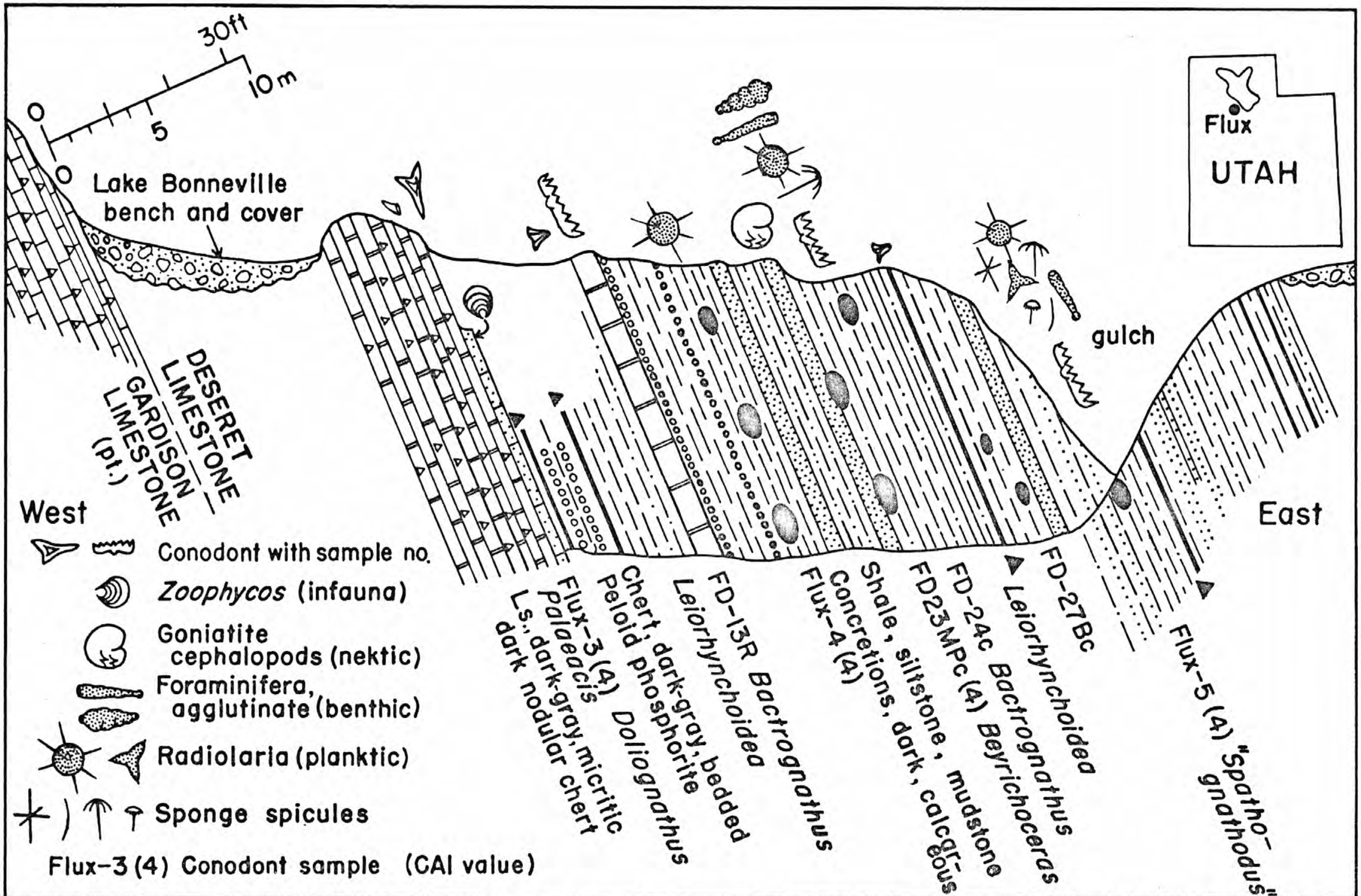


FIGURE 8.--Profile of phosphatic member of Deseret Limestone in exploratory pit and gulch near Flux, Stansbury Mountains. (STOP 5)

phosphorite. The profile of this gulch (fig. 8) permits ready identification of the layers that produced ammonoids, corals, and microfossils. In breaking concretions for ammonoids, care should be taken not to dislodge any of the large calcareous concretions (bullions) that remain in place, as these are valuable for our continuing study of the deep-water origin of the phosphatic member. Conodonts are visible on the upper bedding surface of the massive limestone in the headwall of the gulch. Most of these are ramiform elements, but a few specimens of *Dolignathus latus* may be seen together with small horn corals and the tabulate coral *Paleacis*. Bedding planes in the next higher 0.5 m contain abundant robust *Zoophycos*. Distribution of mainly tiny conodonts in the bullions enclosed in the overlying phosphatic mudstones and shales is erratic. Some bullions are barren, but others contain large numbers of a species of "*Spathognathodus*" of uncertain lineage that may be a homeomorph or descendant of *Bispathodus stabilis* and rare specimens of *Bactrognathus* and *Gnathodus*. All conodonts from this gulch have CAI values of 4. The horizon indicated by conodont sample FD23MPc in the profile (fig. 8) is the horizon of bullions that yielded the ammonoids *Beyrichoceras* and *Djhaprakoceras* studied by Petersen (1969). Bullions at the horizon designated by conodont sample Flux-5 yielded large faunas of planktic spumelline and hagiastrid radiolarians, benthic agglutinate foraminiferans, and sponge spicules, as discussed by Gutschick and Sandberg (1977, 1978). The highest beds on the east side of the gully become increasingly silty upward approaching the contact with the overlying Tetro Member of the Deseret Limestone.

<u>BROAD CANYON (STOP 6)</u>.--This section provides an opportunity to examine the only known succession of parts of the Pinyon Peak Limestone and Leatham Formation, which normally are areally separated, as shown by Sandberg and Poole (1977, fig. 12). The section also contains the most southwesterly known occurrences of the Lodgepole Limestone and its Cottonwood Canyon Member. The profile of the Devonian and Mississippian sequence that can be seen looking northward from just beyond the area of the Indian petroglyphs near the mouth of the canyon is shown on figure 9. Although the Pander Society field trip will not have time to climb to the outcrops, this section merits a return visit because of the excellent exposures of nonresistant beds between the Stansbury Formation and the Paine Member of the Lodgepole. The measured section (fig. 9) is on the east limb of a northward-plunging anticline, as mapped by Palmer (1970). The Stansbury rests unconformably on the Ordovician Fish Haven Dolomite, which is out of sight, just over the crest of the mountain. The contact between quartzitic sandstone of the Stansbury and

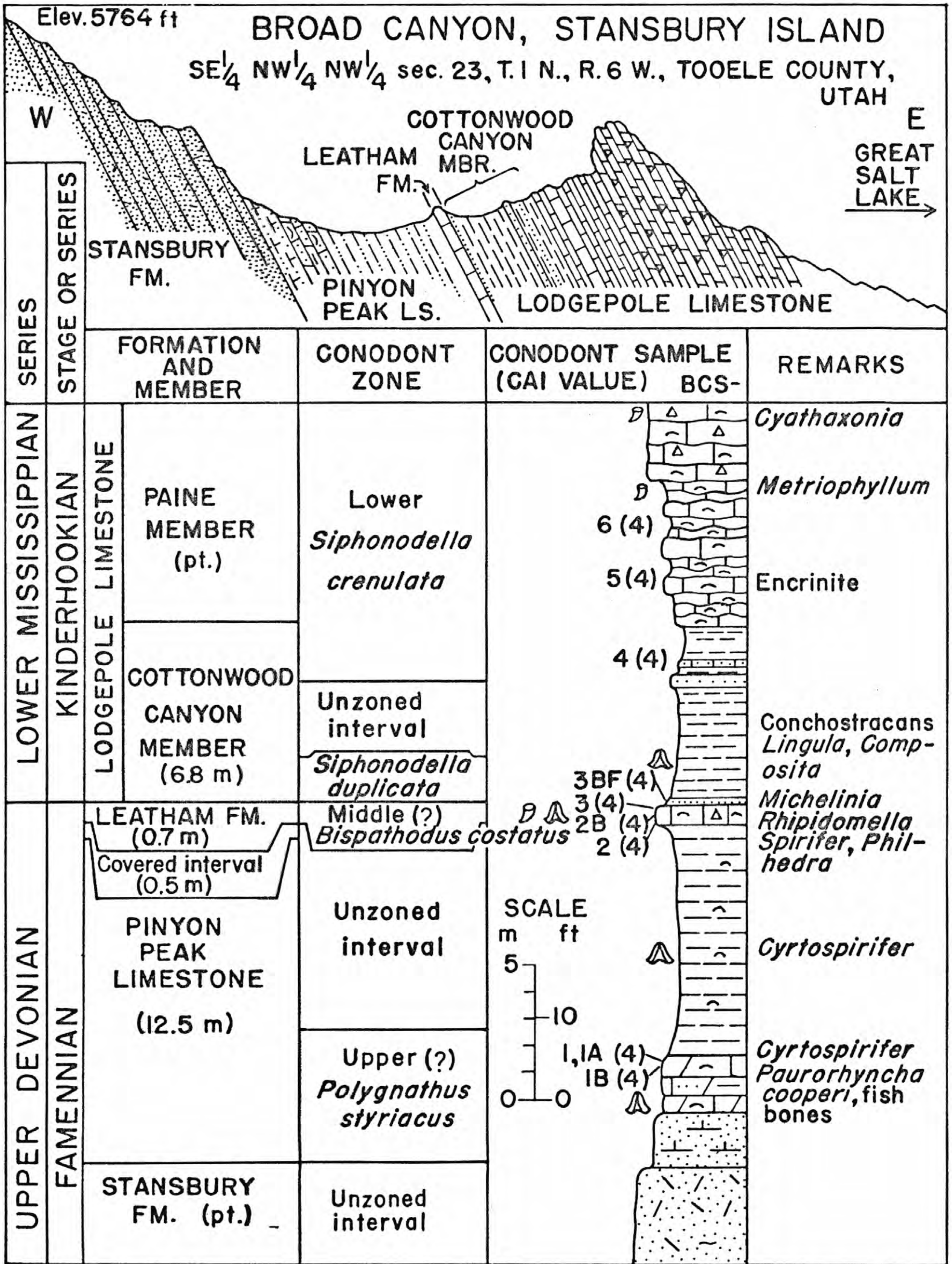


FIGURE 9.--Stratigraphic section and profile of upper part of Stansbury Formation, Pinyon Peak Limestone, Leatham Formation, and lower part of Lodgepole Limestone at Broad Canyon, Stansbury Island. (STOP 6)

fossiliferous calcareous sandstone of the Pinyon Peak is gradational. Conodonts recovered with great difficulty from dolomitic siltstone and partly silicified silty dolomite, 3.7 m above the base of the Pinyon Peak, are tentatively assigned to the polygnathid-icriodid biofacies of the Upper *Polygnathus styriacus* Zone. These and other conodonts from this section have CAI values of 4. The brachiopods *Cyrtospirifer* and *Paurorhyncha* occur in the siltstone and overlying mudstone beds. The next higher bed of dark-gray fossil-fragmental limestone, 0.7 m thick, is correlated with unit 4 of the type Leatham Formation (Sandberg and Gutschick, 1969). It contains silicified megafossils, including small corals and the brachiopods *Philhedra, Rhipidomella, Ambocoelia, Spirifer, Composita,* and *Schizophoria.* The shallow-water conodont fauna consists mainly of long-ranging species, such as *Bispathodus stabilis* and *Polygnathus communis*, but a single specimen of *B. aculeatus* was recovered after processing 10 kg of rock.

The *Siphonodella duplicata* Zone was recognized at the base of the Cottonwood Canyon Member of the Lodgepole Limestone from abundant conodonts on sandy bedding-plane surfaces within a unit of pale-red-purple claystone. Overlying beds of conodont-rich lag sandstone and encrinite contain faunas of the Lower *S. crenulata* Zone. The basal 5.6 m of the overlying Paine Member consists of interbedded encrinite and fossil-fragmental limestone. All beds in this interval yield abundant conodont faunas (200-1,000+ conodonts/kg) of the Lower *S. crenulata* Zone. All species illustrated by Klapper (1966) from the base of the Lodgepole in Montana are represented.

PORCUPINE DAM (STOP 7).--This section on the east limb of the Logan Peak syncline provides one of the few localities where the only Devonian species of *Siphonodella, S. praesulcata,* can be collected. Sandberg, Streel, and Scott (1972, pl. 1, figs. 16-17) illustrated *S. praesulcata* from this locality. Because the roadside exposure is crumbling owing to vehicle traffic and road grading, interested conodont workers are encouraged to collect freely before the tiny outcrop of Leatham Formation is entirely destroyed. In the profile shown in figure 10, the north-south boundary between the Paradise 1:24,000 quadrangle, which was mapped geologically by Mullens and Izett (1964), and the Porcupine Reservoir 1:24,000 quadrangle, which was included on a 1:125,000 geologic map by Williams (1948), is in the lower part of the Little Flat equivalent (member A of Brazer Limestone of Mullens and Izett, 1964). The exposures of the Leatham Formation and lower part of the Lodgepole Limestone are along the road north of the reservoir, beginning 0.25 km east of the spillway. The upper part of the

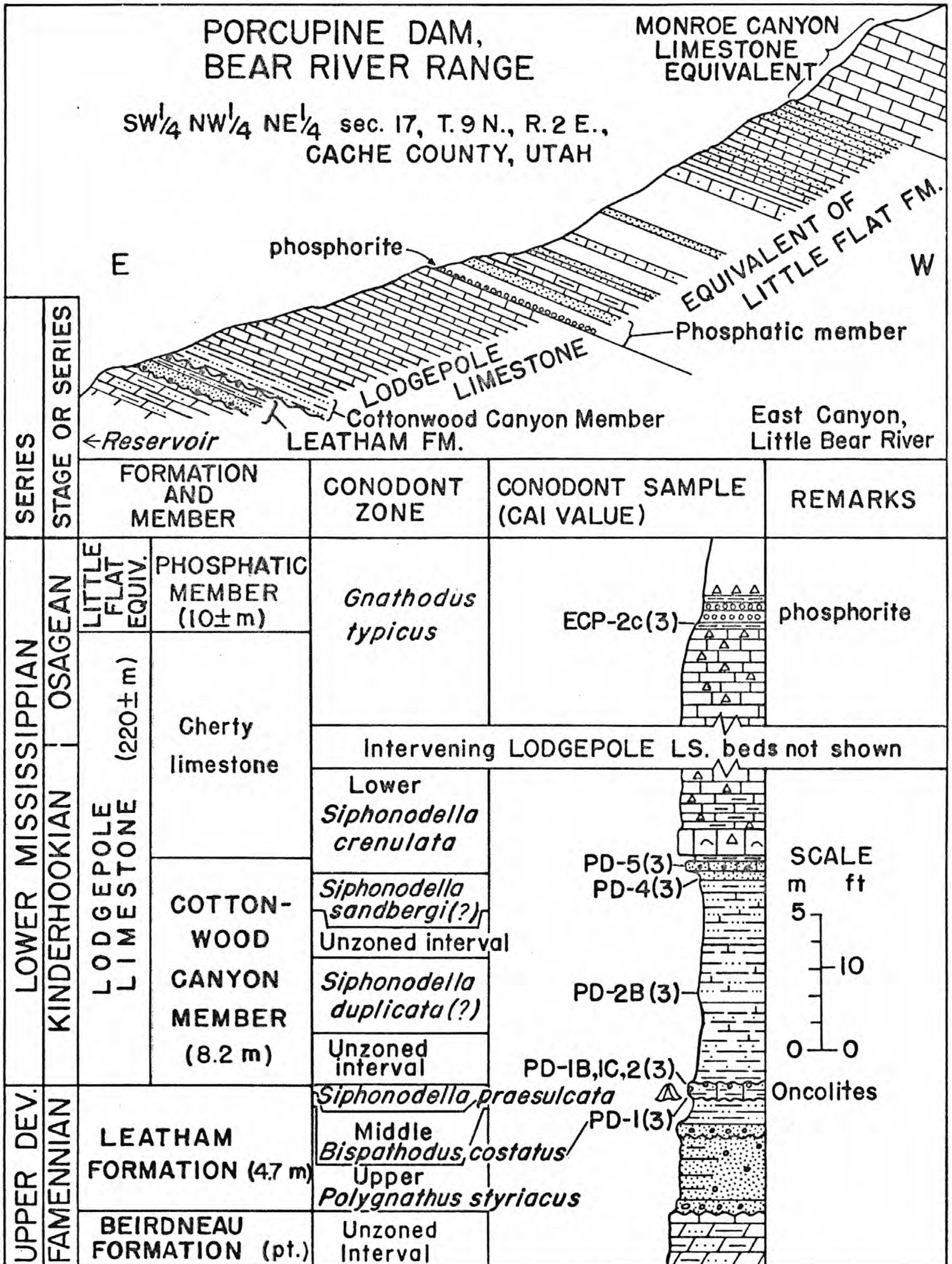


FIGURE 10.--Stratigraphic section and profile of Leatham Formation, Lodgepole Limestone, and adjacent units at Porcupine Dam, Bear River Range. (STOP 7)

Lodgepole and the Little Flat equivalent are best exposed on the hillside south of East Fork of Little Bear River, just west of the spillway.

In the section west of Porcupine dam, the "contact ledge" of Williams (1948) at the top of the Beirdneau Formation has been removed by pre-Leatham erosion. Unit 1 (Sandberg and Gutschick, 1969) of the Leatham Formation consists of conodont-bearing quartzitic conglomeratic sandstone and sandy conglomeratic siltstone, 3.2 m thick. It is overlain by cherty siltstone of unit 3, 1 m thick, and by a ledge, only 0.7 m thick, which combines units 4 and 5. The remainder of unit 5 and all of units 6 and 7 of the Leatham have been removed by erosion beneath the Cottonwood Canyon Member of the Lodgepole Limestone. The lower part of the 0.7-m-thick ledge, at the level of conodont sample PD-1 in the stratigraphic section (fig. 10) yielded a fauna of the Middle *Bispathodus costatus* Zone that includes *B. costatus costatus, B. aculeatus, Palmatolepis rugosa, P. gracilis gracilis,* and *P. gracilis sigmoidalis.* These and other conodonts at Porcupine dam have CAI values of 3. The higher part of this bed (conodont samples PD-1B, 1C, and 2) yielded faunas containing most of these same species together with *Siphonodella praesulcata, Protognathodus meischneri, Polygnathus delicatulus,* and *"Spathognathodus" culminidirectus* (Sandberg and others, 1972, p. 186). Siltstone in the lower 7.7 m of the overlying Cottonwood Canyon contains *Siphonodella* faunas that were recovered with difficulty by crushing. The bed of lag calcarenite (conodont sample PD-5), 25 cm thick, 0.2-0.45 m below the top, however, yielded a huge fauna of the Lower *S. crenulata* Zone. Hand specimens showing conodonts large enough to be visible without a hand lens can be collected from this bed. The basal bed of limestone, 1 m thick, in the overlying undifferentiated part of the Lodgepole contains some visible conodonts and should yield faunas comparable to those from the same stratigraphic position at Broad Canyon.

In the section east and south of the spillway, the contact between the phosphatic member of the Little Flat equivalent and the Lodgepole Limestone has been uncovered by trenching. Conodont sample ECP-2c from a light-gray claystone mixed with encrinite, 15 cm above the base of the phosphatic member, yielded a fauna of the *Gnathodus typicus* Zone. The fauna is dominated by disjunct elements of a *Polygnathus longiposticus* apparatus, but also contains a few specimens of *G. typicus, Pseudopolygnathus multistriatus,* and *Hindeodella segaformis.*

LEATHAM HOLLOW.--On the short drive between Porcupine Dam and Logan, the road crosses Utah Highway 101, which extends eastward into Blacksmith Fork Canyon. The type section of the Leatham Formation, in Leatham Hollow, just north of the

Left Hand Fork, is reached by a drive of only about 14 km from this junction. This section, as photographed by Holland (1952, fig. 9), is well worth visiting. It is one of the few localities where the Leatham, which normally is concealed by talus from the Lodgepole Limestone, is fortuitously exposed. The stratigraphic section of the type Leatham has been illustrated by Sandberg and Poole (1977, fig. 11), who discussed its conodont faunas. The brachiopods, ichnofossils, and epibiont relationships of the Leatham have been described and illustrated by Rodriguez and Gutschick (1967, 1971, 1975).

SPRING HOLLOW (STOP 8).--This section of the Leatham Formation and adjacent units is located 11 km north of the type section and provides an opportunity to examine a nearby and more accessible reference section with only the upper and lower slope-forming units, units 3 and 7, being partly covered. Spring Hollow is on the south side of Logan Canyon and just east of the axis of the Logan Peak syncline. The outcrops are reached by an ascent of about 135 m on the Crimson hiking trail that begins at the group picnic area. The measured section (fig. 11) is located on the north edge of the Logan Peak 1:24,000 quadrangle, but the highway and trail to the group picnic area are on the Mt. Elmer 1:24,000 quadrangle. The area is included on a 1:125,000 geologic map by Williams (1948, pl. 1).

At Spring Hollow, the "contact ledge" of Williams (1948) comprises four units. The basal unit is 0.5 m of laminated, mainly algal-mat limestone. This is overlain by 4 m of cliff-forming pseudobrecciated and pelletal micritic limestone that yielded only fragments of conodonts. The next higher unit of ledge-forming nodular limestone, 3 m thick, was not sampled for conodonts here, but at Leatham Hollow it yielded two conodont faunas of the Middle to Upper *Scaphignathus velifer* Zone. The lower fauna represented the shallower polygnathid-icriodid biofacies, whereas the upper fauna represented a deeper water or more offshore biofacies that contained additionally *Palmatolepis marginifera marginifera* and a new morphotype of *P. distorta*. The highest unit, 2.5 m thick, is yellowish-gray calcareous shale with dark-gray limestone nodules. Brachiopods, mainly *Cyrtospirifer*, may be found in the upper two units.

The Leatham Formation at Spring Hollow contains 6 of the 7 units found at Leatham Hollow. The only missing unit is the silty crinoidal limestone, unit 2, which is 0.7 m thick and occurs just above the basal lag beds at Leatham Hollow (Sandberg and Poole, 1977, fig. 11). The basal 0.25 m of the Leatham, examined in a trench, is carbonaceous shale with limonitic and phosphatic nodules. The overlying unit, which is partly covered, is mainly chertified siltstone and

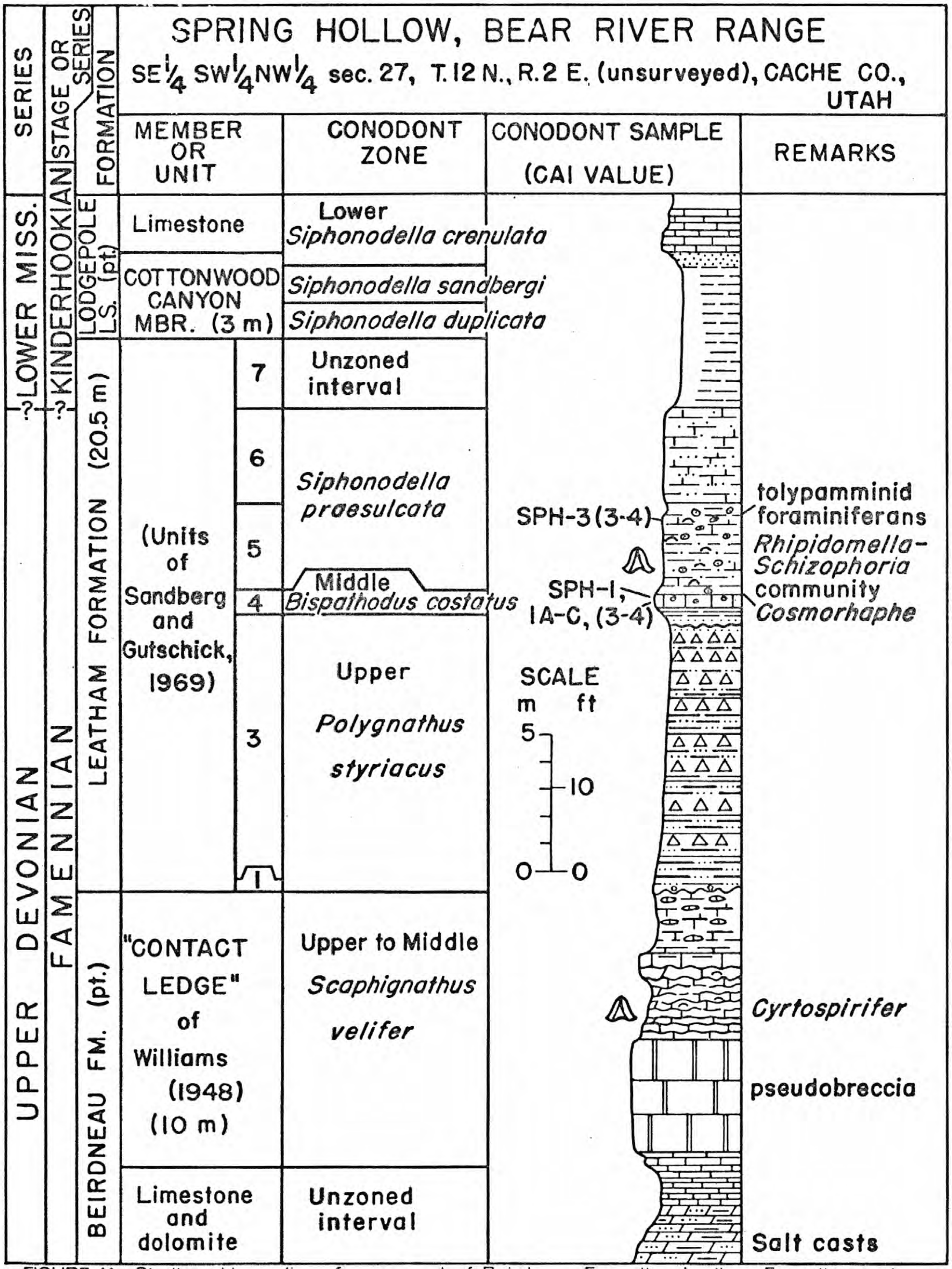


FIGURE 11.--Stratigraphic section of upper part of Beirdneau Formation, Leatham Formation, and lower part of Lodgepole Limestone at Spring Hollow in Logan Canyon, Bear River Range. (STOP 8)

mudstone. The most significant bed in this section is unit 4, which is only 26 cm thick and is gradational upward into the alga-brachiopod biostromal calcareous siltstone and silty limestone of unit 5. The lower 16 cm of unit 4 is dark-gray limestone that contains a few scattered oncolites in its upper part. Conodont samples SPH-1, 1A, 1B, and 1C (fig. 11) from this layer yielded large, diversified faunas of the Middle *Bispathodus costatus* Zone. Unit 5 contains abundant oncolites and brachiopods, which occur free and enclosed in oncolites. The brachiopods are mainly *Schizophoria* (Rodriguez and Gutschick, 1978), *Rhipidomella*, *Syringothyris*, *Spirifer*, and *Beecheria*. The conodont *Siphonodella praesulcata* occurs uncommonly in unit 5.

OLD LAKETOWN CANYON (STOP 9).--This section, which contains excellent exposures of the easternmost thick sequence of the phosphatic member of the Little Flat Formation, is located on the north side of Utah Highway 30, 1.5 km east of Laketown at the south end of Bear Lake (fig. 1). The area is included in the geologic map of the Randolph quadrangle (Richardson, 1941). The succession from the top of the Beirdneau Formation through the Pennsylvanian part of the Wells Formation was shown in a stratigraphic section by Sando, Dutro, and Gere (1959, fig. 5). Their Mississippian units A, B, and C constitute the Little Flat Formation, 244 m thick, and their units D, E, and F, the overlying Monroe Canyon Limestone, 85 m thick. A stratigraphic section of the phosphatic member of the Little Flat Formation was shown by Sando, Dutro, Sandberg, and Mamet (1976, fig. 2). A profile and stratigraphic section of the phosphatic member, giving the position of conodont samples, are shown on figure 12. The conodonts from the phosphatic member have CAI values of 1½, whereas those from the underlying Lodgepole Limestone have CAI values of 1½-2. On the basis of low temperatures of 50°-140°C (Epstein and others, 1977) suggested by these values, the area around Bear Lake was predicted to be one of the most favorable areas for petroleum exploration of the Cordilleran hingeline in Utah (Sandberg and Gutschick, 1977a, p. 9). A gas well, the American Quasar No. 1-20 Hogback Ridge, was subsequently completed in the Permian Phosphoria Formation and Triassic Dinwoody Formation, about 10.5 km northeast of the Old Laketown Canyon section.

The strata in the Old Laketown Canyon section are nearly vertical, and measurement was complicated by bedding-plane faults that locally cut abruptly across beds from one horizon to another. Several duplicate sections had to be measured and compared to ensure the accuracy of the composite stratigraphic section shown on figure 12. The conodont faunas from beds of encrinite in the

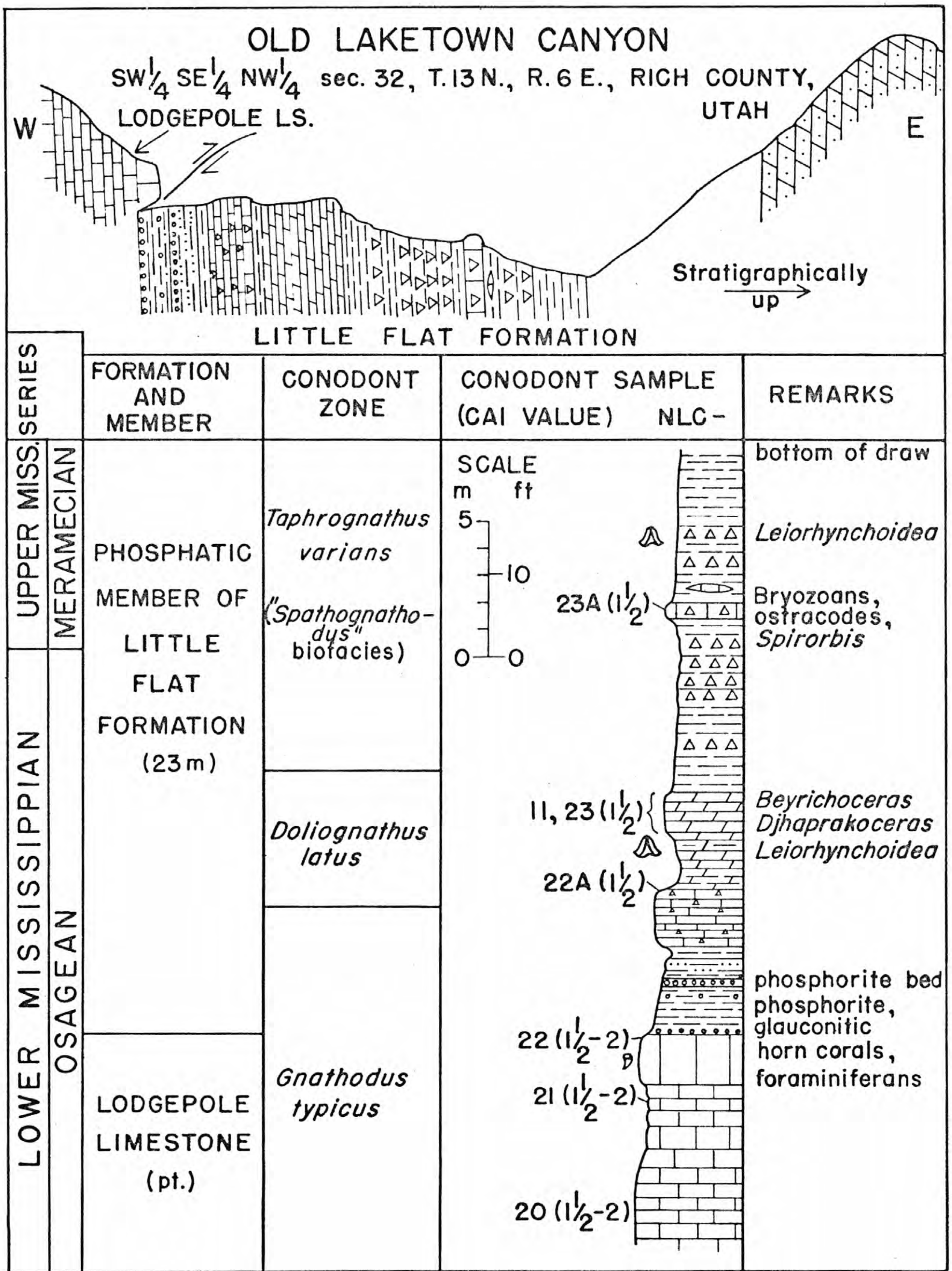


FIGURE 12.--Stratigraphic section and profile of upper part of Lodgepole Limestone and phosphatic member of Little Flat Formation at Old Laketown Canyon. (STOP 9)

upper part of the Lodgepole Limestone are dominated by *Polygnathus communis communis*, but also contain *Pseudopolygnathus* cf. *P. multistriatus*, *Anchignathodus penescitulus*, *"Spathognathodus" elongatus*, and *Pelekysgnathus bultyncki*. Collections from carbonate interbeds in the lower part of the Little Flat Formation contain *Doliognathus latus* as well as a possible ancestral species.

BRAZER CANYON.--The sheer fault escarpment on the west side of the Crawford Mountains from Rex Peak, the highest peak on the south, to the mouth of Brazer Canyon on the north is formed largely by the Brazer Dolomite, 260 m thick. This entire sequence, deposited on the prograding carbonate shelf, is equivalent in age to the basinal phosphatic member of the Little Flat Formation, 23 m thick, just 22 km to the northwest at Old Laketown Canyon (fig. 1). Although the difference in thickness of time-equivalent sediments between the two localities is about 11:1, this does not necessitate an escarpment for the shelf margin as steep and impressive as the present fault escarpment of the Crawford Mountains. Assuming that the present post-faulting separation of 20 km across the depositional strike between the two localities is roughly equal to their original depositional separation, that the shelf margin was exactly midway between them, and that the pre-depositional seafloor was virtually flat, a slope of only 2° would account for the difference in thickness of the sediments that had accumulated by early Meramecian time. Allowing a 50 percent margin for error in these assumptions, we infer a slope of $2° \pm 1°$ for the slope between shelf margin and basin. Only if the original slope of the seafloor were as great as 2°-3°, could the resulting slope have been as great as 5°.

The type section of the Brazer Dolomite, which is located in a strike belt parallel to and 1.2-2 km east of the mine road in Brazer Canyon, and three nearby sections on the same strike belt were measured and described by Sando, Dutro, and Gere (1959). They also mapped the geology of the area between and adjoining the four measured sections. The units and lithologies of the type Brazer are summarized in the Stratigraphic Glossary. Unfortunately, time will not permit the climb up a side canyon of Brazer Canyon to examine the type section from top to base. On a return visit, however, it would be possible to approach from the south and east to within a short distance of the base of the section by use of a four-wheel-drive vehicle.

WARNER HOLLOW (STOP 10).--Time, weather, and road conditions permitting, the Pander Society field trip may visit Warner Hollow, 5.5 km south of the mouth of Brazer Canyon. Under favorable conditions, a four-wheel-drive vehicle can be

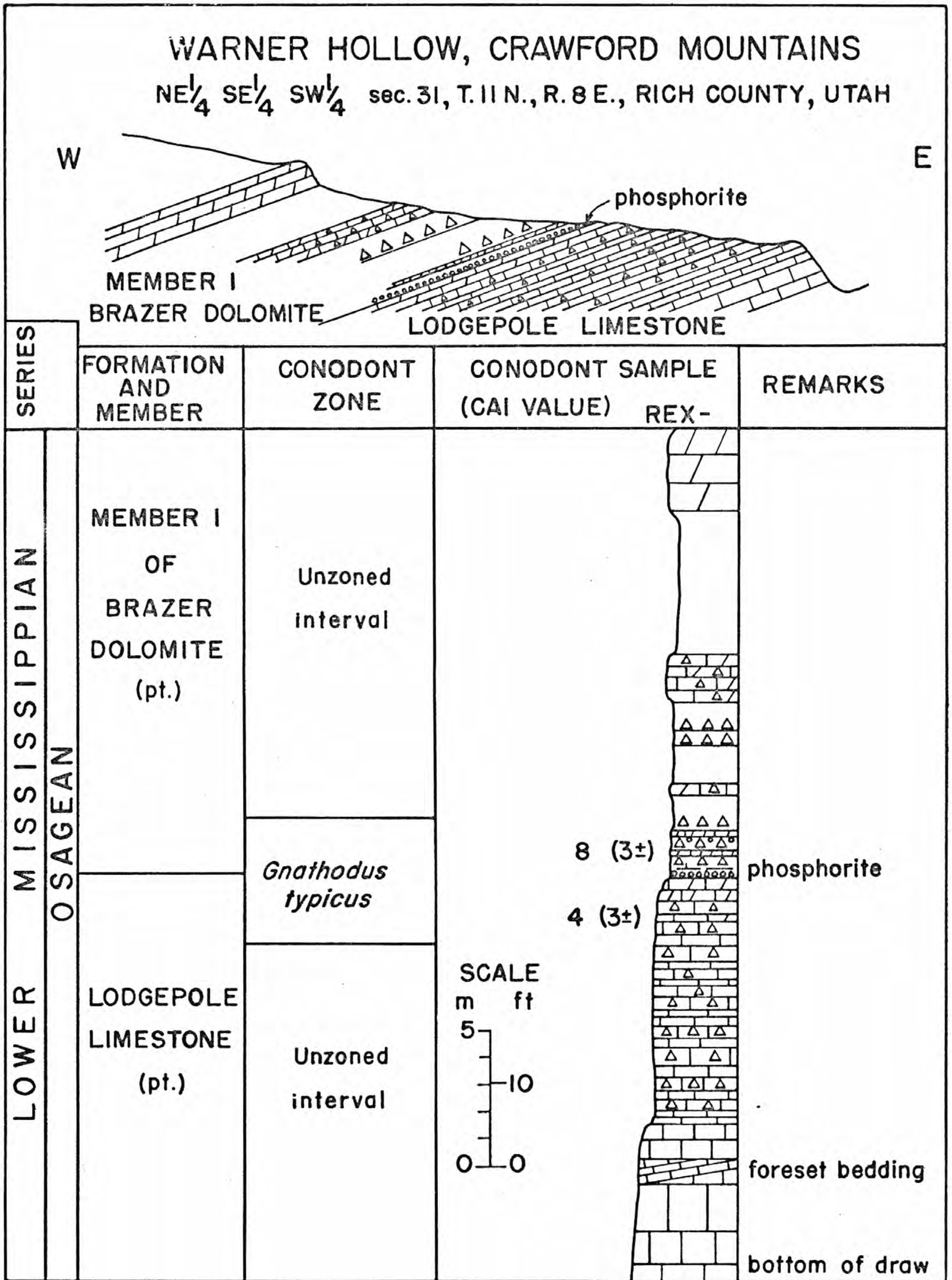


FIGURE 13.--Stratigraphic section and profile of upper part of Lodgepole Limestone and lower part of Brazer Dolomite at Warner Hollow, Crawford Mountains. (STOP 10)

driven to within 100 m of this locality. Warner Hollow provides an opportunity to visit the easternmost occurrence of the phosphatic interval, which is at or just above the base of the Brazer Dolomite, depending on the criteria used to define the contact with the underlying Lodgepole Limestone. The equivalent stratigraphic interval is, unfortunately, concealed in the type section of the Brazer. The Warner Hollow locality is included on a geologic map by Sando, Dutro, and Gere (1959, fig. 1); it is on the west side of the major drainage, about 0.7 km south of their section 4. The stratigraphic section and profile of this locality are shown on figure 13. The exposed basal ∿2 m of the phosphatic interval comprises in ascending order: pelletal phosphorite, 13 cm thick; interbedded yellowish-gray phosphatic mudstone and cherty dolomitic limestone, 0.7 m thick; and interbedded yellowish-gray calcitic argillaceous dolomite and black chert, ∿1.1 m thick. The thin basal ledge of this upper unit, indicated by conodont sample REX-8 on figure 13, yielded a large fauna (500 conodonts/kg) of the *Gnathodus typicus* Zone. This fauna is dominated by *Polygnathus communis communis, Gnathodus typicus, Pseudopolygnathus nudus,* and *Hindeodella segaformis.* These and all other conodonts recovered from the Lodgepole and Brazer at four localities in the Crawford Mountains, regardless of whether or not they came from close to the phosphatic interval, have an unusual pale-yellowish-brown matte color that does not match any of the colors in the CAI scale (Epstein and others, 1977, fig. 5). Because the conodonts still retain white matter and have a definite light-brown tint, they are assigned a CAI value of ∿3, even though their surface texture and grayness might indicate a somewhat higher value.

BIOSTRATIGRAPHIC, SEDIMENTOLOGIC, AND PALEOGEOGRAPHIC INTERPRETATIONS

The interpretations for Famennian, Kinderhookian, and Osagean and Meramecian rocks are largely summarized and updated from more comprehensive papers by Sandberg and Poole (1977), Sandberg and Klapper (1967), and Sandberg and Gutschick (1977a), respectively. The interpretations of rates of sedimentation have not appeared elsewhere. The sedimentational-faunal and conodont biofacies models (figs. 15, 16) were previously presented as slides in a talk (Sandberg and Gutschick, 1977b) on April 28, 1977 to the Pander Society Symposium on Carboniferous Conodonts at the meeting of the North-Central Section of the Geological Society of America at Carbondale, Illinois.

Famennian rocks

The "contact ledge" of Williams (1948) represents a segment of a middle Famennian depositional complex that was deposited during the time of the *Scaphignathus velifer* Zone (Sandberg and Poole, 1977, p. 163-166). Rocks of this complex also include the Costigan Member of the Palliser Formation in the Alberta Rocky Mountains, the Trident Member of the Three Forks Formation in western Montana and east-central Idaho, and possibly the lower member of the Pinyon Peak Limestone in southwestern Utah and adjacent parts of Nevada and Arizona. In the Western United States this complex is narrowly restricted to a northward-trending belt on both sides of the Sevier thrust system (Sandberg and Poole, 1977, fig. 10). The middle Famennian complex was deposited in a long, narrow, shallow seaway along the former continental shelf-cratonic platform boundary of the older Upper Devonian depositional complex. This landward shift of the seaway was in response to a pulse of the Antler orogeny, as a result of which most of eastern Nevada was emergent and undergoing erosion. This complex contains the well-known *Cyrtospirifer monticola* brachiopod fauna and *Platyclymenia (Pleuroclymenia) americana* ammonoid fauna. The conodont faunas (Klapper and others, 1971) represent the polygnathid-icriodid and polygnathid-pelekysgnathid biofacies of the Middle to Upper *Scaphignathus velifer* Zone.

The time-equivalent upper member of the Pinyon Peak Limestone (including the lowermost few meters of the overlying Fitchville Formation) and the basal beds (units 1-3) of the Leatham Formation were deposited during the time of the *Polygnathus styriacus* Zone on the outer cratonic platform and on the continental shelf and slope, respectively, of a late Famennian seaway (Sandberg, 1976, fig. 2). The Pinyon Peak contains conodont faunas assigned to the polygnathid-icriodid as well as the clydagnathid biofacies. The clydagnathid biofacies was found recently at City Creek Canyon, 13 km northeast of Salt Lake City (fig. 1). Previously, this biofacies had been known only from western Colorado and north-central Wyoming (Sandberg, 1976). The basal beds of the Leatham contain conodont faunas assigned to the palmatolepid-polygnathid biofacies and, west of the report area, to the more offshore palmatolepid-bispathodid biofacies. The upper member of the Pinyon Peak, lowermost beds of the Fitchville, and basal beds of the Leatham are included in the upper Famennian depositional complex (Sandberg and Poole, 1977, p. 166-172) that extends from northern Montana southward into eastern Nevada, and then southeastward into southeastern Arizona and southwestern New

Mexico. The seaway in which this complex was deposited had moved farther eastward than the middle Famennian seaway and had inundated a wide expanse of the cratonic platform in response to a major pulse of the Antler orogeny.

The Devonian lower part of the Fitchville Formation, which is 34 m thick at the type locality and 9 m thick at Rock Canyon and Flux (fig. 2), represents mainly the polygnathid-icriodid biofacies of the Lower *Bispathodus costatus* Zone, except for the lowermost beds, which represent the highest part of the underlying Upper *Polygnathus styriacus* Zone. In addition to conodonts, it contains abundant clisiophyllid and caninoid corals that are comparable to those in the Etroeungt Limestone of Belgium. This part of the Fitchville is included in the uppermost Famennian depositional complex (Sandberg and Poole, 1977, p. 172-176), although its areal distribution is different from that of younger beds of the complex (units 4-7 of the Leatham Formation and their equivalents). In western Wyoming, strata representing the Lower *B. costatus* Zone were assigned to the lower tongue of the Cottonwood Canyon Member of the Madison Limestone at most localities and to the undifferentiated lower part of the Madison at Dinwoody Canyon and Windy Gap by Sandberg and Klapper (1967), who could not determine a western connection for this first transgression by the Madison sea. The dating of the Devonian lower part of the Fitchville, together with its lithology and distribution, suggests that it was once depositionally continuous with the Devonian lower part of the Madison in western Wyoming. The seaway in which these time-equivalent beds were deposited extended northeastward from the southwest corner of Utah through western Wyoming into Montana in approximately the same area as the seaway in which the upper member of the Pinyon Peak Limestone and equivalents had been deposited. The connection between the Utah and Wyoming segments of the seaway was apparently across the west end of the Uinta Mountains (fig. 1). Evidence for this connection has been largely destroyed by pre-Mississippian erosion there, although it is possible that small uneroded remnants of rocks representing the Lower *B. costatus* Zone will be found.

Units 4-7 of the Leatham Formation, which represent the highest Devonian Middle *Bispathodus costatus* and *Siphonodella praesulcata* Zones and include at the top a few meters of undated beds that might be Mississippian (fig. 3), are assigned to the uppermost Famennian depositional complex (Sandberg and Poole, 1977, p. 172-176). These units of the Leatham form a short segment of a remarkably uniform sequence of beds that was deposited in a narrow, originally continuous belt from northern Montana to southern Nevada. Lithogenetically, sequentially, and faunally identical strata are assigned to the Sappington Member of the Three Forks Formation in Montana and to the middle member of the Pilot Shale in western Utah and eastern Nevada. This Sappington-Leatham-middle Pilot depositional subcomplex consists, in ascending order, of a basal lag sandstone; a conchostracan- and *Tasmanites*-bearing mudstone; an alga-brachiopod biostromal (oncolitic) silty limestone; and calcareous siltstone and sandstone interbedded locally with silty mudstone, carbonaceous shale, and oolitic limestone. The subcomplex was deposited in a great variety of very shallow water environments including mud flats, lagoons, algal banks, and offshore bars. Unit 4 of the Leatham, because of its common *Palmatolepis*, represents the only offshore, open-marine environment. Conodonts are generally scarce and difficult to recover by conventional methods, except at a few localities such as Porcupine Dam (fig. 10), Spring Hollow (fig. 11), and Lick Creek Road, Montana (Sandberg and others, 1972). Locally, conodonts have been obtained by crushing the basal lag sandstone (Sandberg, *in* Klapper and others, 1971, fig. 5, col. 1). Brachiopods, sponges, agglutinate foraminiferans, and ichnofossils have been described from the subcomplex by Rodriguez and Gutschick (1967, 1971, 1975), Gutschick (1962), Gutschick and Perry (1959), and Gutschick and Lamborn (1975). The upper part of the uppermost Famennian depositional complex, which represents the Middle *B. costatus* and *S. praesulcata* Zones, lies west of the lower part of the complex, which represents the Lower *B. costatus* Zone. Thus the Sappington-Leatham-middle Pilot depositional subcomplex constitutes the first record of a widespread Famennian regression after the time of deposition of the Logan Gulch Member of the Three Forks and pre-"contact ledge" beds of the Beirdneau Formation. This subcomplex was regressive in relation to the former western Wyoming seaway but transgressive in relation to the eastern shore of a low landmass that bordered the Antler highlands to the west. Consequently, the latest Devonian was a time of diminished Antler orogenic activity.

Kinderhookian rocks

The Cottonwood Canyon Member of the Lodgepole Limestone in Utah, which is continuous with the upper tongue of the member in western Wyoming (Sandberg and Klapper, 1967), contains a sequence of conodont faunas assigned to the Kinderhookian *Siphonodella duplicata, S. sandbergi,* and Lower *S. crenulata* Zones. Although the earliest Kinderhookian *S. sulcata* Zone may also be present locally, it has not yet been identified. The *S. sulcata* Zone is present, however, farther south in

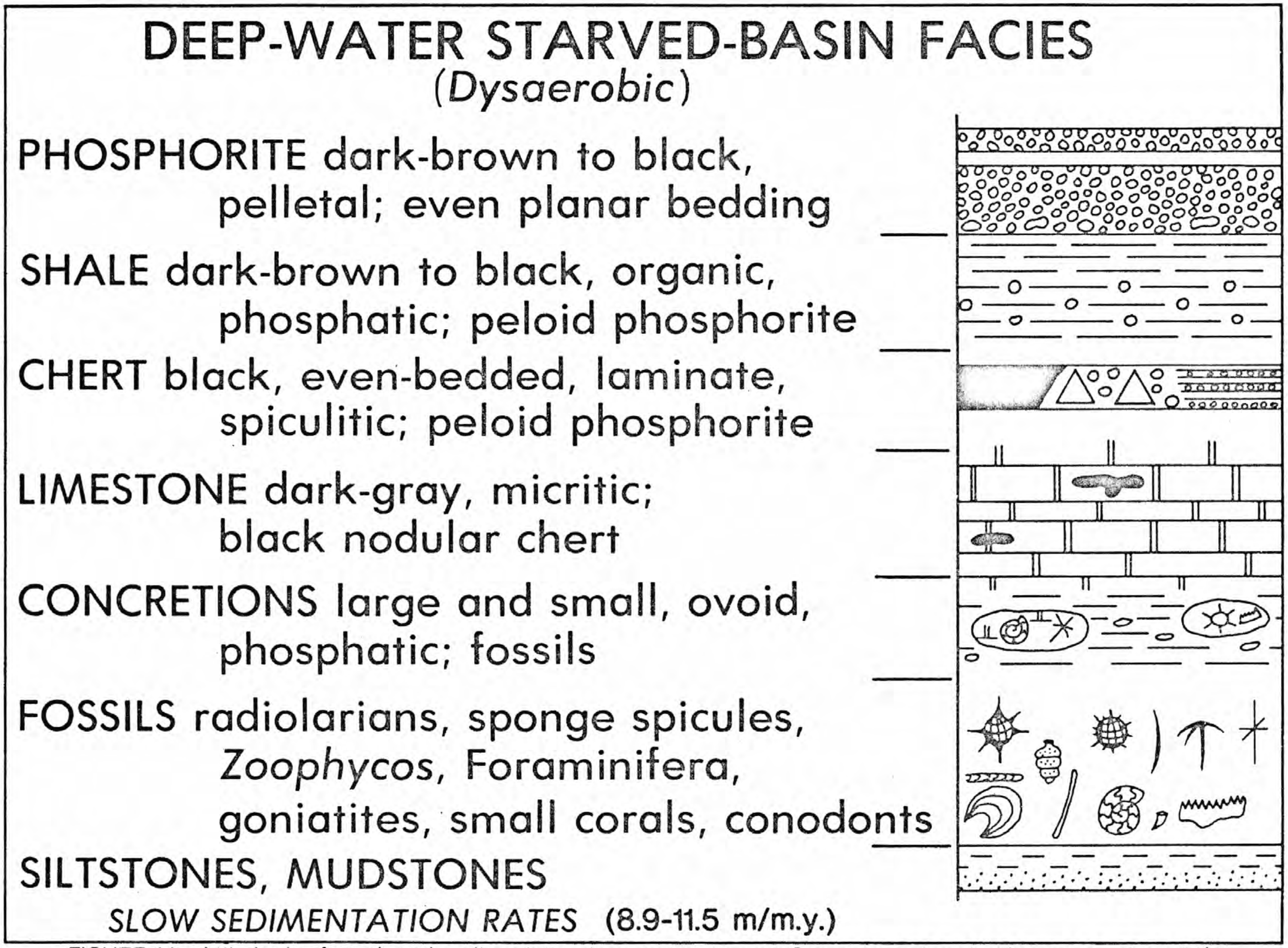
Utah in equivalent beds at the base of the middle part of the type Fitchville Formation (Sandberg and Poole, 1977, fig. 13) in the East Tintic Mountains (fig. 1). In Utah, the Cottonwood Canyon, which overlies a major regional unconformity, was laid down in an open-marine environment of a northeastward-transgressive sea. Because of a slow rate of sedimentation, its conodont faunas represent winnowed lag concentrates of conodonts of virtually the same age. Reworking of conodonts from unconformably underlying rocks is relatively minor, and most identifiable reworked conodonts were derived from the Upper *Polygnathus styriacus* Zone. Megafossils, chiefly brachiopods, have been found in the mudstone, siltstone, and shale between lag beds, together with ichnofossils such as *Zoophycos*. The basal beds of the overlying Paine Member of the Lodgepole and undifferentiated equivalents are transgressive glauconitic encrinites and fossil-fragmental nodular limestones that are related to and depositionally continuous with the Cottonwood Canyon Member; they also contain conodont faunas of the Lower *S. crenulata* Zone.

Osagean and Meramecian rocks

The phosphatic member of the Deseret Limestone and equivalents was deposited during Osagean and early Meramecian time in deep water (∿300 m) of a sediment-starved basin, which occupied the eastern half of a foreland basin that was bordered on the west by the Antler orogenic highlands. The elongate starved basin, the axis of which is now oriented about N. 25° E. in western Utah, was about 160 km wide and more than 400 km long. The basin was bordered on the east by a westward-prograding carbonate platform or bank, which was made up of a thick wedge of bioclastic sediments interpreted as a "stratigraphic reef" by Rose (1976). The topographic relief of the foreslope of this "stratigraphic reef" was interpreted to be perhaps 200-400 m by Rose (1976, p. 458). Our reevaluation of the pertinent biostratigraphic and sedimentologic data, in collaboration with W. J. Sando, is in agreement with these figures. However, we interpret that the seaward face of the shelf margin in Utah was not steep, like that of the Guadalupian "reef" of southeastern New Mexico, but rather had a gentle foreslope of 2° $\pm$ 1°, or at most 5°, as calculated under the heading, Brazer Canyon, between the shelf margin and basin. Regardless of its local relief, the carbonate bank served as a sediment trap that effectively diminished but did not totally restrict the westward transport of craton-derived terrigenous sediments into the basin. A flysch trough that occupied the western part of the foreland basin received most of the coarse clastic sediments eroded from the emergent Antler highlands farther west. At first, only suspensions of clay-size particles filtered through the flysch-trough sediment trap, but by early Meramecian time, as the flysch trough was becoming filled, reworked silt-size distal flysch sediments spread eastward into the western part of the starved basin and intertongued with the phosphatic mudstones and phosphorites.

The phosphatic member in the central part of the starved basin comprises mainly dark-colored beds of pelletal, peloidal, and oolitic phosphorite; organic phosphatic shale enclosing large calcareous concretions that were termed bullions by Holdsworth (1966); bedded black spiculitic and radiolarian chert (Lydite of German usage); cherty micritic limestone; and carbonaceous siltstone and mudstone. All these sediments were slowly deposited at rates of 8.9-11.5 m/m.y., as calculated under the heading, Rates of Sedimentation. The characteristics that influenced our interpretation of a deep-water starved-basin facies are summarized on figure 14. It should be emphasized that no single sedimentologic or faunal characteristic is diagnostic of a deep-water origin. Considered in total, however, these characteristics, together with the absence of the shelly fauna (large colonial and horn corals, bryozoans, gastropods, and spiriferid and productid brachiopods) that characterized equivalent strata of the carbonate platform and the absence of extensive bioturbation *(Cosmorhaphe)* that characterizes well-oxygenated distal flysch sediments, mandate a water depth of ∿300 m. Originally, we interpreted the starved-basin facies to be euxinic (Sandberg and Gutschick, 1977a), anoxic, or anaerobic. However, after reconsidering our finding of benthic organisms (sponges, agglutinate foraminiferans, small horn corals, and the brachiopod *Leiorhynchoidea carbonifera*) in the bullions and the scarcity of pyrite in the bullions and other sediments, we now interpret a dysaerobic or minimal-oxygen setting for the deeper part of the starved basin. A dysaerobic or reduced-oxygen zone was defined by Rhoads and Morse (1971) to contain 1.0 to 0.1 ml of dissolved oxygen per liter of sea water. Besides the benthic fauna, we have recently found the deep-water variety of *Zoophycos* infaunal traces in cherts and limestones. The basinal sediments also contain abundant planktic radiolarians, listed by Gutschick and Sandberg (1978), and nektic goniatites and conodonts that fell to the sea floor after death.

On the eastern periphery of the starved basin, the phosphatic member contains four types of transported sediments that are interbedded with the indigenous planar-bedded, clinoform, cherty slope limestones, as shown diagrammatically in



FIGUFE 14. --Lithologic, faunal, and sedimentational characteristics of Osagean and early Meramecian deep-water starved-basin facies..

our starved-basin sedimentational and faunal model (fig. 15). The first type of introduced sediments are crossbedded encrinite and calcarenite debris flows, which are shown in surface and cross-section views at the right of the model. The debris flows, which are related to Waulsortian-type mounds, are composed mainly of debris from crinoid thickets on the upper slope and shelf margin. The small spheres on the slope at the left of the model represent light-colored phosphorite oolites, which form on the slope and then are transported basinward by currents. We interpret the light coloring to result from chemical precipitation in an aerobic setting. These phosphorite oolites commonly are found mixed with the encrinite and calcarenite debris flows as well as with the dark-colored amorphous peloids and pellets of phosphorite that formed in the dysaerobic setting of the central basin. Some of the light-colored oolites may be banded or encased in dark amorphous phosphorite, suggesting their post-transport enlargement under dysaerobic conditions. A third type of transported sediment, shown on the left, is bedded chert that we believe originally formed on the slope and then periodically slid basinward, as indicated by arrows. Such chert has contorted or recumbently folded bedding, in contrast to the planar-bedded basinal chert shown in the front of the model. The fourth type of transported sediment is craton-derived sandstone and sandy carbonate, which periodically was transported through channels in the shelf margin and deposited in the basin. The distal-slope sediments of the phosphatic basin contain fragmented shallow-water bryozoan (indicated by # in the model) and echinoderm debris, infaunal traces of *Cosmorhaphe* and *Scalarituba,* benthic hyperamminid and saccamminid foraminiferans, and nektic as well as nektobenthic conodonts.

The biofacies model that we postulate for early and middle Osagean conodont faunas in Utah is shown by figure 16. At the left of this model with a greatly exaggerated vertical scale are shown the positions of the photic, aerobic, and dysaerobic zones, which are separated by heavy dashed lines. The boundaries between conodont living-zones or niches are shown by thinner, smaller dashed lines. The model is based on analysis of 22 collections from 12 localities, totalling about 3,700 conodonts. The largest and most diversified collection comes from a mixed encrinite-oolitic phosphorite debris flow. The diversity of this fauna, which includes at least 17 platform species, and its abundance (500+ conodonts/kg) suggest that the slope provided the optimum environment for the growth and preservation of conodonts, possibly because of food provided by basinal upwelling.

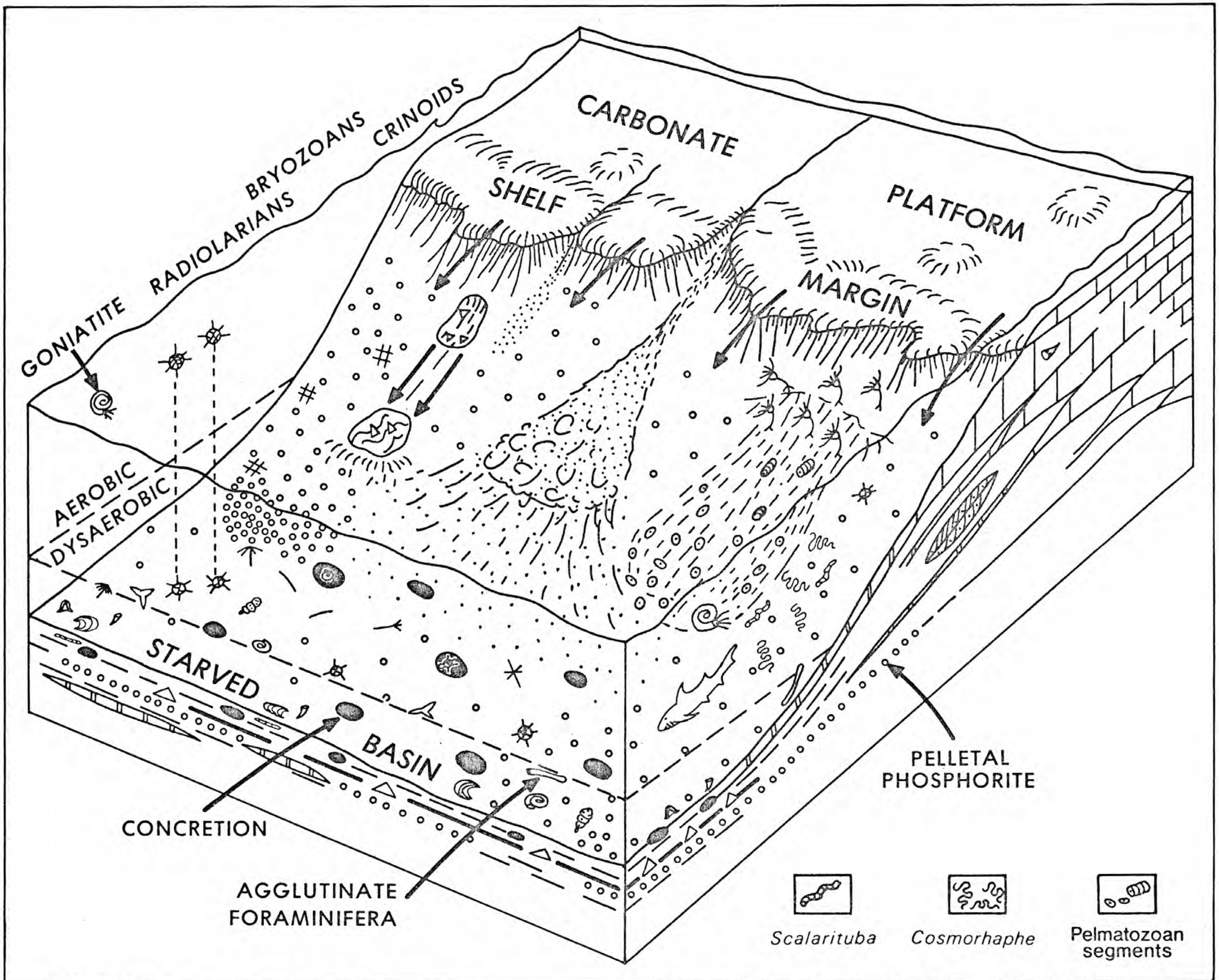


FIGURE 15.--Sedimentational and faunal model for Osagean and early Meramecian starved basin and adjacent carbonate platform. Vertical scale greatly exaggerated; actual slope 2° ± 1°.

Of the counted conodonts, 65 percent are *Polygnathus communis communis*. This aptly named subspecies is ubiquitous. Consequently, we interpret that it lived at a shallow depth in the photic zone. Nearly as widespread, but not as abundant, are two species of spathognathodiform conodonts, which each account for 6 percent of the faunas. One of these is a new species of *"Spathognathodus"* that may be a homeomorph or descendant of *Bispathodus stabilis*. The other is *Anchignathodus penescitulus*. Both species probably lived in the shallow, photic zone, but *A. penescitulus*, which is more abundant landward, favored a nearshore setting above the foreslope, margin, and platform, whereas *"S."* n. sp., which is more abundant seaward, favored a setting farther from shore. Faunas in the most offshore setting at the left of the model are composed almost exclusively of *"S."* n. sp. and *P. communis communis*.

Perhaps our most interesting interpretation is that *Gnathodus* and *Pseudopolygnathus*, each making up about 7 percent of the total counted faunas, lived in the aerobic zone and were probably nektobenthic. The interpretation is based on the abundance of these genera in platform, margin, and slope faunas that were recovered either from light-colored limestones or from dark limestones that contain light-colored and presumably more oxygenated phosphorite. Only one or two stragglers of either genus have been found in dysaerobic dark limestones containing dark phosphorite or in bullions. But why do we consider these genera nektobenthic? Because if *Gnathodus* or *Pseudopolygnathus* had lived any distance above the bottom, then some post-mortem fallout into basinal sediments would have occurred as in the case of *Doliognathus*, *Scaliognathus*, and *Bactrognathus*.

All other genera in the faunas are found in smaller quantities. *Doliognathus* and *Scaliognathus*, each composing about 1 percent of the counted faunas, apparently lived in the middle of the aerobic zone, because they occur farther seaward than does *Bactrognathus*, which also composes about 1 percent of the faunas but is scarce in bullions. *Bactrognathus*, *Gnathodus*, *Pseudopolygnathus*, *Doliognathus*, and *Scaliognathus*, as indicated by their initials on figure 16, also ranged into shallower water of the platform margin, but their shoreward extent into the backslope or platform setting in the Uinta Mountains has not yet been analyzed. *Eotaphrus burlingtonensis* occurs mainly in the shallow-water setting of the carbonate margin and upper foreslope. Although it constitutes well under 1 percent of the counted faunas, it composes 1 to 18 percent of the collections in which it occurs. Not shown in the model, but occuring in 6 collections from the starved basin, slope, and margin, are small numbers of *Hindeodella segaformis*. *Polygnathus communis carina*, which is abundant in early Osagean faunas from the Upper Mississippi Valley,

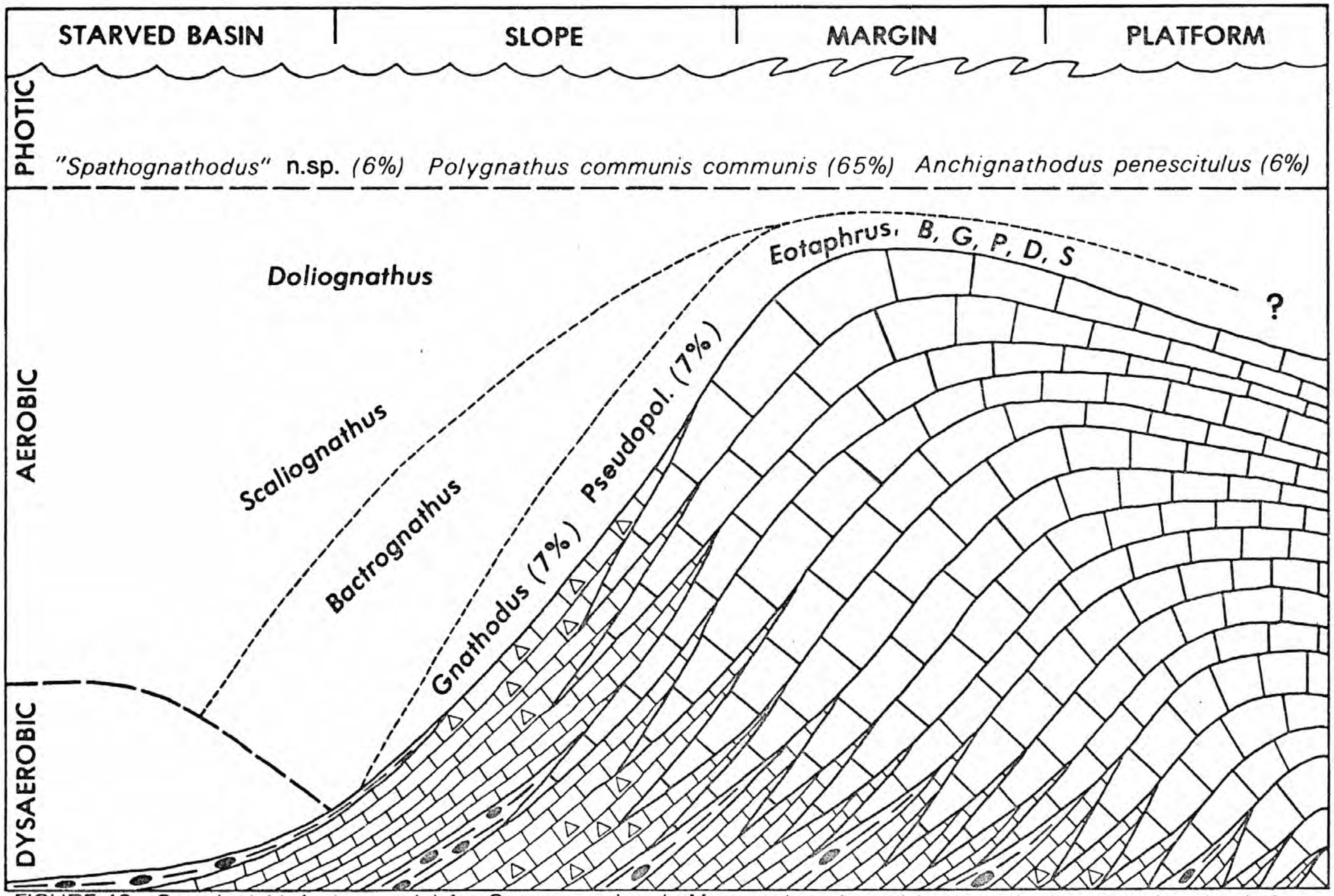


FIGURE 16.--Conodont biofacies model for Osagean and early Meramecian starved basin and adjacent carbonate platform. Vertical scale greatly exaggerated; actual slope 2° ± 1°.

is extremely rare in basinal faunas, because it apparently lived in very shallow water of the margin and platform.

The model postulated for late Osagean and early Meramecian faunas of the basin and platform is similar to that shown by figure 16, except that most platform genera had died out by then. Only "*Spathognathodus*" n. sp. lived on in the starved-basin setting, where it occurs together with the newly evolved *Taphrognathus varians*, whereas in the aerobic slope and shelf settings, *Gnathodus* cf. *G. texanus* (or *G. linguiformis*) occurs with these two species.

Our interpretation of conodont biofacies serves as a working model, which should be useful to students of other Osagean carbonate-platform margins. It also may be applicable to other regions as well as to conodont faunas of other ages.

Rates of sedimentation

The rates of sedimentation of the phosphatic member of the Deseret Limestone and equivalents, the Cottonwood Canyon Member of the Lodgepole Limestone and equivalents, and the thick carbonate interval between these two thin intervals can be calculated by using the average length of a conodont zone as a yardstick. The rate of sedimentation of carbonate rocks of the shelf margin cannot be calculated as accurately, because their sparse conodont faunas are not amenable to precise conodont zonation. However, their rate is believed to be comparable to the more readily calculable rates of the lithologically and sedimentologically similar Gardison Limestone or Woodhurst Member of the Lodgepole, which underlie the phosphatic member in different areas. By substituting rates for these units, it should be possible to contrast the rates of sedimentation for the phosphatic member and the laterally equivalent shelf carbonate rocks. However, as stated by Sandberg and Poole (1977, p. 176), who applied this method of using conodont zones as a yardstick to the Upper Devonian Pilot Shale sediments in Utah and to Lower Mississippian Antler flysch sediments in Idaho, "it should be remembered that the rates are not exact, because they are calculated assuming the same order of magnitutude, not the same length of time, for the duration of...conodont zones." Furthermore, the length or timespan of a single Paleozoic conodont zone is too short to either be confirmed or discredited by radiometric age determinations or by zonations based on other organisms.

The average length of an Upper Devonian conodont zone was estimated to be one-half million years (0.5 m.y.) and the length of the Kinderhookian *Siphonodella isosticha* Zone (approximately equivalent to the present *S. isosticha*-Upper *S. crenulata* Zone) to be about 2-3 m.y. by Sandberg and Poole (1977). Independently, Sando (1977, p. 495) arrived at figures of 1.3 and 1.5 m.y., respectively, for the average timespans of Lower and Upper Mississippian conodont zones. Sando's calculation was based on Collinson, Rexroad, and Thompson (1971), who used the zonation of Collinson, Scott, and Rexroad (1962) for the type Mississippian of the Upper Mississippi Valley area. The Kinderhookian was divided into 5 zones and the Osagean (to the top of the Keokuk Limestone) into 4 zones by Collinson, Scott, and Rexroad (1962, chart 6), for a total of 9 Lower Mississippian conodont zones. Although the zonal names differ, this same total of 9 zones is recognized in the Western United States: 6 Kinderhookian *Siphonodella* zones (Sandberg, Ziegler, and others, 1978) and 3 Osagean zones (Sandberg, 1978). Thus, Sando's figure of 1.3 m.y. would be applicable to the Western United States. We believe because of sedimentological considerations that the *S. isosticha*-Upper *S. crenulata* Zone is about twice the length of other *Siphonodella* zones. Thus, if the 2-3 m.y. figure of Sandberg and Poole (1977) for this zone were halved, then the resulting 1-1.5 m.y. figure for any other *Siphonodella* zone would bracket Sando's 1.3 m.y. figure for an average Lower Mississippian conodont zone. Thus, the independent figures of Sando (1977) and Sandberg and Poole (1977) appear to be in close agreement. Considering that we are dealing with an order of magnitude and not with absolute time, we believe it is reasonable to use a timespan of 2-3 m.y. for the *S. isosticha*-Upper *S. crenulata* Zone and 1.3 m.y. for all other Lower Mississippian conodont zones in calculating the rates of sedimentation of Lower Mississippian rocks in Utah.

The time during which the phosphatic member of the Deseret Limestone was deposited in the central part of the Deseret basin is estimated to span one-quarter of the *Gnathodus typicus* Zone, all of the *Doliognathus latus* Zone, and one-quarter to one-half of the Osagean lower part of the *Taphrognathus varians* Zone. This part eventually may be defined as a separate formal zone. This estimate gives a total of 1.5 to 1.75 whole conodont zones of 1.3 m.y. timespan, or a rounded total of 2-2.3 m.y. The phosphatic member in the central basin (e.g., at Broad Canyon and Flux) averages 54 m in thickness, but includes terrigenous clastic sediments interbedded in its upper half. Using the above figures, an average rate of sedimentation for the phosphatic member in the central basin is calculated to be 23.5-27 m/m.y. A more realistic rate for starved-basin phosphatic shale sedimentation alone, excluding introduced clastics, and eliminating the necessity of estimating parts of zones, can be calculated by using the thickness of the *Doliognathus latus* Zone, which lies entirely within the phosphatic member. At Buckhorn Canyon in the Dugway Range (fig. 1), this zone is represented by 11.5 m of shale, lying between 3.5 and 15 m above the base of the phosphatic member of the Woodman Formation. The rate for sedimentation of this starved-basin phosphatic shale is calculated to be 8.9 m/m.y. This rate compares closely with the 4.5-6 m/m.y. rate calculated by Sandberg and Poole (1977) for the starved sedimentation during the early history of the Late Devonian Pilot basin in western Utah and eastern Nevada. Using a timespan of 2-2.3 m.y., the rate of sedimentation of the entire phosphatic member, which is 23 m thick and contains little interbedded clastics, at Old Laketown Canyon is calculated to be 10-11.5 m/m.y. This rate is close to the rate for the *Doliognathus latus* Zone at Buckhorn Canyon. The phosphatic member at Rock Canyon is ∿60 m thick and is mainly two slope-deposited limestones bounded by shale and siltstone (figs. 2 and 6). Using the same timespan, its rate of sedimentation is calculated to be 26-30 m/m.y. Thus the calculations show that within the phosphatic member slope limestone was deposited nearly three times as rapidly as basinal phosphatic shale (26-30 vs. 8.9 m/m.y.)

The rate of sedimentation of the Cottonwood Canyon Member of the Lodgepole Limestone, which consists mainly of transgressive lag (or condensed) deposits, as described by Sandberg and Klapper (1967, p. B29), was exceedingly slow. At Leatham Hollow and Broad Canyon (figs. 3, 2), two Kinderhookian *Siphonodella* zones, representing a timespan of 2.6 m.y., are condensed into an interval of 3-6.8 m or less, if allowance is made for part of the Lower *Siphonodella crenulata* Zone in the upper part of the interval. Thus the rate of sedimentation of the member must have been only about 1-3 m/m.y.

Compared to the rates of sedimentation of the phosphatic member and Cottonwood Canyon Member, the rate of sedimentation of the thick intervening carbonate interval was extremely rapid. The Paine and Woodhurst Members of the Lodgepole Limestone and the equivalent upper part of the Fitchville Formation and the Gardison Limestone were deposited during the time of a small part (arbitrarily estimated as one-quarter) of the Lower *Siphonodella crenulata* Zone (1/4 x 1.3 m.y.), all of the *S. isosticha*-Upper *S. crenulata* Zone (2.3 m.y.), and an estimated one-half to three-quarters of the overlying *Gnathodus typicus* Zone (1/2-3/4 x 1.3 m.y.) This sequence has a thickness range of about 183 to 309 m in Utah (figs. 2, 3). Applying a total timespan of 3-4.3 m.y. to a thickness of 183-309 m, the range of sedimentation rates for the entire intervening carbonate interval is calculated to be 43-103 m/m.y.

Deposition of the upper part of the carbonate interval, comprising the equivalent Gardison Limestone and Woodhurst Member of the Lodgepole Limestone, was even more rapid than that of the interval as a whole. Between Broad and Rock Canyon (fig. 2), the Gardison and equivalents have a thickness range of about 200-275 m and an average thickness of about 240 m. Dominantly coarse bioclastic limestones in the interval were deposited during the estimated three-quarters of the *Gnathodus typicus* Zone that is not represented by the lower part of the phosphatic member of the Deseret Limestone. Using a timespan of 1 m.y. (3/4 x 1.3 m.y.), the rate of sedimentation of the Gardison is calculated to have a range of 200-275 m/m.y. and an average of about 240 m/m.y. This rate is nearly as rapid as the rate of 267-400 m/m.y. calculated for Early Mississippian Antler flysch sedimentation in Idaho (Sandberg and Poole, 1977).

We believe that a rate comparable to 240 m/m.y. is the maximum rate applicable to sedimentation of shelf carbonate rocks such as the Brazer Dolomite in the Crawford Mountains (fig. 3). Using this figure, their rate of sedimentation was about 10 times that of the phosphatic member in the central basin (240 m/m.y. vs. 23.5-27 m/m.y.). Another means of estimating the rate of sedimentation of shelf carbonate rocks is to compare the 260-m-thick Brazer Dolomite at Brazer Canyon with the 23-m-thick phosphatic member at Old Laketown Canyon. As previously calculated, the rate of sedimentation of the phosphatic member there is 10-11.5 m/m.y. If the timespans of the two sequences were identical, then a simple proportion would indicate that the Brazer Dolomite, which is 11.3 times as thick, was deposited at a rate of 113-130 m/m.y. The true rate of sedimentation of shelf carbonate rocks probably lies within the extremes of 113 to 240 m/m.y. Whichever set of calculations is used, however, this rate was at least 10-11 times that of equivalent basin sediments.

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