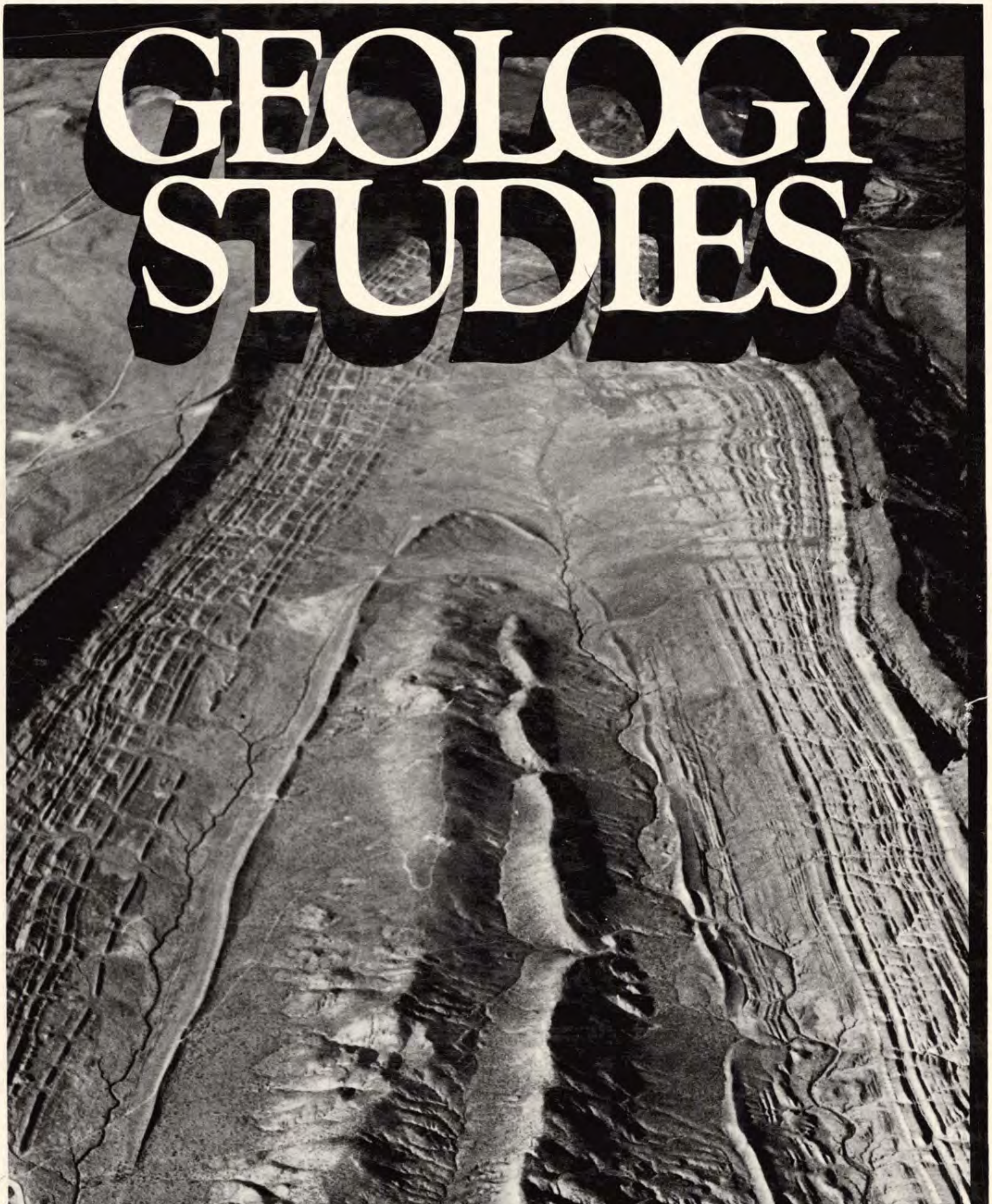
BRIGHAM YOUNG UNIVERSITY

GEOLOGY STUDIES



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Cover: Virgin anticline near St. George, Washington County, Utah.

Petrology and Petrography of the Great Blue Formation at Wellsville Mountain, Utah*

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ABSTRACT.—Great Blue Formation (upper Mississippian) at the Deweyville section of Wellsville Mountain is composed of seven distinct facies. Three members are formed by these seven facies: (1) lower limestone member; (2) median shale member, and (3) upper limestone member. Facies forming each member are as follows:

upper limestone member
7. mudstone-packstone dolomite sandstone facies
6. wackestone facies

median shale member
5. mudstone-wackestone claystone facies

lower limestone member
4. mudstone-packstone facies
3. oolite facies
2. wackestone-packstone facies
1. wackestone dolomite sandstone facies

These facies accumulated in an unstable depocenter in the eastern portion of the Cordilleran miogeosyncline and are interpreted to have formed in the following environments: (1) shoaling ooid-producing marine waters; (2) very small localized reefs or banks of mostly solitary corals, accompanied by bryozoans and crinoids forming behind the shoal; (3) skeletal detritus scattered behind the small localized reefs or banks of skeletal development, debris being transported cratonward by tidal action, storms, and local currents; (4) a pelletal enriched zone farther cratonward behind the zone of skeletal detritus; and (5) lagoonal conditions behind the pelletal zone and nearest to the craton, all forming in zones Y and Z as described by Irwin (1965). Each facies has many microshifts within itself to produce subfacies. Shifts or oscillations of the environment of deposition within these facies may have been abrupt, producing sharp contacts between individual units with occasional complete changes in rock type. A well-developed diastem was located within the uppermost mudstone packstone dolomite sandstone facies and may represent how sharp contacts between individual units may have formed. Reducing conditions may have been active behind the shoal and small localized reefs or banks. Chert occurs sporadically within the section and represents disproportionate amounts of silica entering the depocenter in time. Silica making up chert sequences may have been supplied by streams draining the craton or from pulsations within the Antler orogenic belt.

INTRODUCTION

Wellsville Mountain is located in central northern Utah, the northernmost extension of the Wasatch Mountains (fig. 1). Elevation of the area is 1,310 meters at the valley floor and 2,860 meters at Box Elder Peak directly east of Honeyville. Rocks at Wellsville Mountain range from Precambrian to Pennsylvanian in age and in general dip northeast at approximately 30-40 degrees. The west face of Wellsville Mountain offers excellent, well-exposed, but steep areas for field investigation. Canyons along the mountain front are steep and narrow. There are no permanent trails on the mountainside.

Accessibility to the study area is from Utah Highway 69 via gravel roads which lead up to and parallel the mountain front. The immediate study area is accessible only on foot (fig. 2). The terrain is very rugged in the principal study area, with steep slopes and occasional cliff-forming sequences being common.

Previous Work

Wellsville Mountain was first mapped by Beus (1958) and Gelnett (1958), each mapping half the mountain—north and south halves, respectively. Prior to 1958, geologists studying the area described only portions of the mountain. Williams (1943) studying the Carboniferous of the northern Wasatch Mountains, applied the name Brazer Limestone to limestones of Late Mississippian age in the Dry Lake area, 15 kilometers southeast of the study area, dividing the formation into 5 units. Williams and Yolton (1945) studied both Brazer and Lower Wells formations at Dry Lake and Deweyville sections, adding units 6 and 7 in the Lower Wells Formation to the pre-

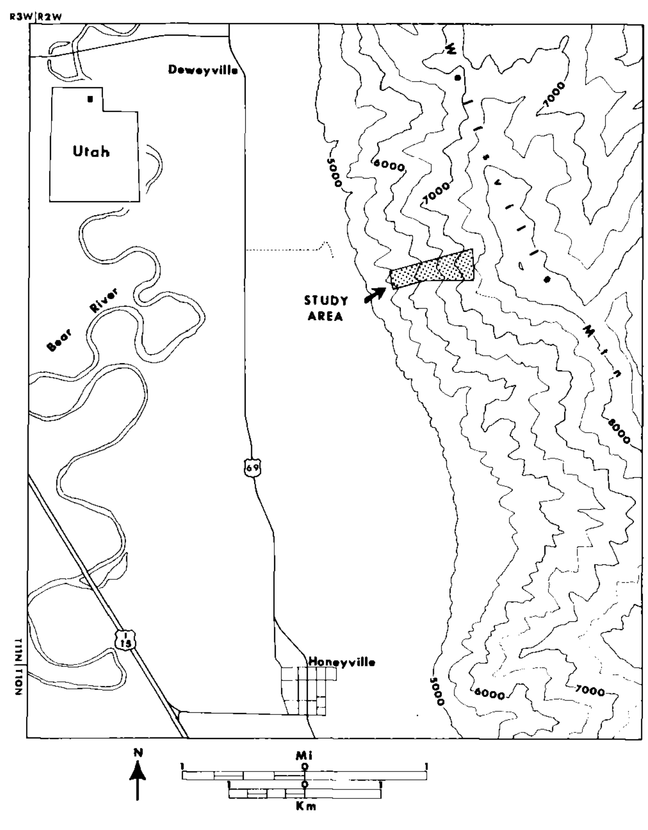


FIGURE 1.—*Index map of study area.*

*A thesis presented to the Department of Geology, Brigham Young University, in partial fulfillment of the requirements for the degree, Master of Science, December 1976; Harold J. Bissell, thesis chairman.

FIGURE 2.—West face of Wellsville Mountain south of Deweyville, Utah, with route of ascent and study area south of Holdaway Canyon outlined.

vious 5 Brazer Formation units. Parks (1951) made the first detailed study of corals within the Brazer Limestone at Dry Lake section. Sadlick (1955) studied the Mississippian-Pennsylvanian boundary at Dry Lake section and divided unit 1 of the Brazer into the Humbug Formation, units 2, 3, and 4 into lower limestone, Long Trail Shale, and upper limestone of the Great Blue Limestone, respectively, and unit 5 into Manning Canyon Shale. Nygreen (1958) later renamed Wells Formation the Oquirrh Formation. Williams (1958) in the *Geologic Atlas of Utah, Cache County,* agreed with Sadlick (1955) that units 1-5 of Brazer Limestone could be subdivided as follows: Unit 1, Humbug Formation; unit 2, lower limestone; unit 3, Long Trail Shale; and unit 4, upper limestone of the Great Blue Limestone; and unit 5, Manning Canyon Shale. Crittenden (1959) used Dry Lake section of Brazer Limestone in his comparison of Mississippian stratigraphy in both allochthonous and autochthonous areas of the central Wasatch and western Uinta Mountains. Carr and Trimble (1961) used Dry Lake section in applying the name Great Blue Formation to the Deep Creek Mountains of southern Idaho. Butkus (1975), in a regional investigation of the Great Blue Limestone, studied sections at Dry Lake and Logan Canyon, as well as sections in central Utah. Allan Swede (graduate student, Utah State University) is currently studying corals of the Great Blue Formation from Dry Lake section eastward into the Logan Canyon area. As used herein, the name Great Blue Formation will be applied, following the usage of Morris and Lovering (1961), who modified the terminology of Gilluly (1932), who used the name Great Blue Limestone.

Statement of the Problem

The purpose of this paper is to determine by petrologic and petrographic studies the environment of deposition of the Great Blue Formation at Wellsville Mountain south of Deweyville, Utah. Attention was directed to both detailed field and laboratory studies of a carefully measured section not only to determine the conditions of sedimentation of the exposed Great Blue Formation, but to form a working model of the ancient sedimentary environment. Rose (1976, p. 465), in his study of Mississipian carbonate shelf margins of the western United States, pointed out the need for "detailed field investigations in appropriately located mountain ranges . . . to compare such features with other ancient counterparts, as well as with features in different parts of the same Mississippian shelf."

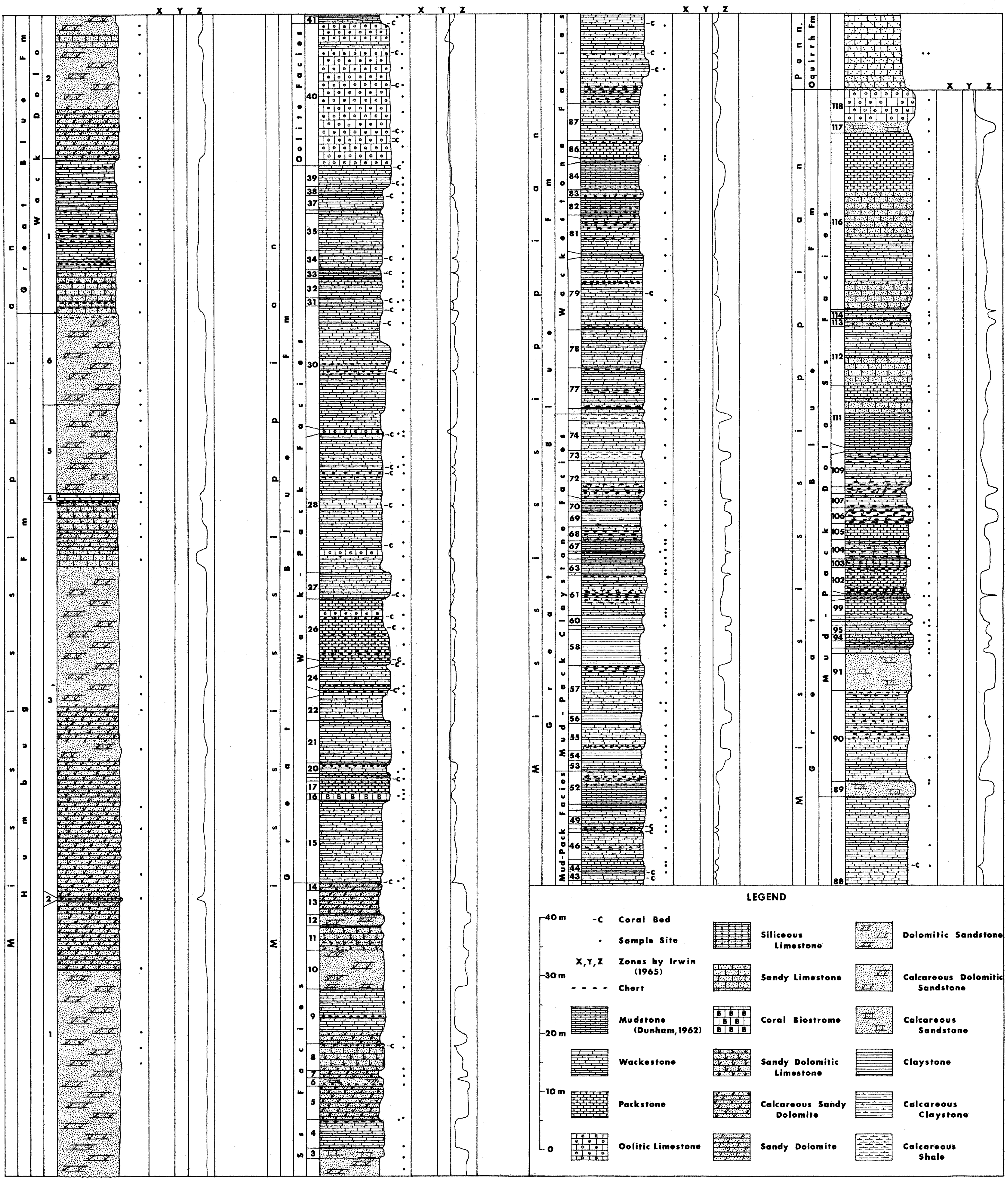
ACKNOWLEDGMENTS

The writer wishes to thank Dr. Harold J. Bissell, who suggested the problem and served as thesis chairman, and whose guidance, suggestions, and petrographic help were always present. Dr. James L. Baer is to be thanked for serving as committee member. Dr. J. Keith Rigby identified sponge spicules and bryozoans and provided helpful suggestions. Dr. William J. Sando offered suggestions helpful in constructing the exact problem to be studied. The John S. Berge Fund provided for field and laboratory expenses sustained during this study. Mr. Ralph Stegen and Mr. Scott McMullin were of great help as field assistants. I admire and appreciate my wife, Debbie, for her unending support of this thesis project.

METHODS

Methods of study of the Great Blue Formation were divided into field and laboratory studies. Field studies were conducted from July through October 1975, with brief revisiting in August 1976. Location of the measured section is the ridge south of Holdaway Canyon, approximately one-third to two-thirds of the way up the mountainside, in the SW$\frac{1}{4}$, sec. 15, T. 11 N, R. 2 W, directly east of the 12-mile marker on Utah Highway 69. Preliminary measurement of the section was accomplished by the use of a five-foot Jacob staff, Abney hand level, and Brunton compass; bright yellow spray paint was applied to the section at five-foot intervals, and every twenty-foot interval was numbered. Detailed field descriptions, following preliminary measurements of the section, were made. Field classification of carbonate rocks used the classification scheme of Dunham (1962). Final measurement of the section was accomplished using a 100-foot steel tape and a Brunton compass as a check of preliminary measurements; data accrued by steel tape measurements were corrected to obtain true thickness. Through the measured section beginning at the base of Humbug Formation through the Great Blue Formation into lowermost Oquirrh Formation, 248 samples were collected. Exact locations of collection sites are labeled on the stratigraphic column (fig. 3). Rock color was determined by comparison with the rock-color chart distributed by the Geological Society of America (1970). Photography of important field locations was carried on in conjunction with sampling of the measured section, and I obtained oblique airphotographs in July 1976. Using both 1"x2" and 2"x3" petrographic slides, I prepared 248 thin sections. Those suspected to contain dolomite were stained with Alizarin Red S to detect true amounts of dolomite and calcite (Friedman 1959). Thin sections were described by complete scanning of each one to learn relative amounts and types of (1) skeletal, pelletal, intraclastic, oolitic, and clastic admixtures; (2) matrix; (3) recrystallization or other diagenetic effects; (4) energy which affected the ancient sea bottom; and (5) environment in which the sediments were deposited, insofar as it could be pinpointed. Textural types and skeletal fragments were identified by the use of excellent carbonate texts by Bissell (1970),

FIGURE 3.—Stratigraphic column of measured Deweyville section of Humbug and Great Blue formations.



Carozzi and Textoris (1967), Horowitz and Potter (1971), and Majewske (1969). No one classification was used to name or describe the various carbonate rocks in thin-section. Dunham's classification (1962) served well for a base name with the classification of Leighton and Pendexter (1962) serving as a modifier in front of the base name.

Corals were studied by using standard sectioning and acetate peel methods. Acetate peels actually reveal more corallite structure than direct observations of original cut surfaces, thereby facilitating coral identification. No detailed coral study was attempted because the research has been done by Parks (1951), Butkus (1975), and currently the corals are being studied by Allan Swede (graduate student, Utah State Univ.). Identified corals were used for lateral correlation with Dry Lake section, where all three of the above-mentioned geologists made investigations.

GEOLOGIC SETTING

Great Blue Formation at Wellsville Mountain was deposited in the subsiding Great Blue basin, Late Mississippian forerunner to the Oquirrh basin. Great Blue basin subsidence and sediment infilling formed in the easternmost part of the Cordilleran miogeosyncline. Uplift in Late Devonian time began to affect the eastern portion of the miogeocline and was centered around the Stansbury Mountains, southwest of Wellsville Mountain, forming the Stansbury uplift (Rigby 1959, Morris and Lovering 1961). Lower Mississippian time was a transgressive phase over beveled and eroded uplifted areas which deposited the Lodgepole Limestone. Uppermost Lodgepole Limestone may be an eastward thinning section of Deseret Limestone. Broad epierogeny after Lodgepole or Deseret Limestone deposition formed an unconformity which now separates lower Mississippian sequences from upper Mississippian sequences (Beus 1958; Sando 1967, 1968, 1975). Epierogeny gave way to slow subsidence as Humbug Formation was deposited with further deepening of the basin depositing Great Blue Formation. Mississippian-Pennsylvanian boundary at Dry Lake section is within the Manning Canyon Shale (Sadlick 1955). Northwest of Dry Lake section to my measured section reveals no Manning Canyon Shale to be present; accordingly, Oquirrh Formation rests unconformably on the Great Blue Formation (Beus 1958). This latest Mississippian through earliest Pennsylvanian unconformity forms a wedge which has progressively northward left a thinner section preserved (Beus pers. comm. 1975; Williams 1943, 1958; Williams and Yolton 1945; Mullens and Izett 1964).

Elsewhere, the Antler orogeny in Late Devonian into Mississippian time began to affect central Nevada, forming an arcuate belt which extended northerly through central Nevada into parts of central Idaho. Poole (1974) cites deformation by eastward compression to have affected the Antler orogenic belt, forming the Antler foreland basin, and possibly warping the continental crust under the shelf causing broad epierogeny and basin subsidence farther eastward into the miogeosyncline. Compressive force was from partial closure of a marginal ocean basin west of the Antler orogenic belt due to active subduction at an island arc near the Klamath Mountains, California (Burchfiel and Davis 1972, 1975).

Wellsville Mountain now occupies the northern portion of the Wasatch Front (also more accurately referred to as the Wasatch hinge line). In Cretaceous time the hinge line served as an eastern terminus where thrusts carrying large slabs of crust came to rest after translating allochthonous sections eastward. Wellsville Mountain represents part of the allochthonous section thrust upon the hinge line and is related to the Willard thrust to the southeast (Crittenden *et al.* 1971). This large allochthonous plate in northern Utah is referred to as the Cache allochthon by Crittenden (1972). Amount of displacement of the Cache allochthon may have been as much as 64 to 160 kilometers in northern Utah (Crittenden [1961], pers. comm. Sept. 1976), with locations east of Ogden, Utah, where the thrust overlies other smaller thrusts (Eriksson 1960, Rigo 1968, Armstrong 1968). Evidence would then suggest that Great Blue Formation at Wellsville Mountain was originally deposited 64 to 160 kilometers west of its present position within the Cordilleran miogeosyncline and not near the hinge line and shelf sequences east of the hinge line. Total eastward translation of the allochthonous plate could have been as little as 11-16 kilometers as suggested by Brady (1965) and Black (1965) for the Mount Nebo thrust (of approximate similar age) 221 kilometers farther south.

Most recent diastrophism to affect Wellsville Mountain was vertical uplift by the Wasatch fault which raised the mountain to its present elevation. Initial uplift along the Wasatch fault began in Miocene time and has continued to present, forming the Wasatch Mountains and the easternmost terminus of the Great Basin.

STRATIGRAPHY

Upper Mississippian stratigraphy at Wellsville Mountain and adjacent areas has undergone revision of formation names. Mississippian limestone units in the Crawford Mountains were named Brazer Limestone by Richardson (1913), from Brazer Canyon 9.6 kilometers northeast of Randolph, Utah. Fossils were scarce, but G. H. Girty (in Richardson 1913) assigned Brazer Limestone a lower upper Mississippian age. No type locality or reference section was designated for the Brazer Limestone by Richardson (1913). Later, in mapping the geology and mineral resources of the Randolph quadrangle, Richardson (1941) again did not designate type or reference sections for the Brazer Limestone. Brazer Limestone was then correlated by Mansfield (1927) into southeastern Idaho. Later authors then applied the name Brazer Limestone to upper Mississippian limestone sequences in which interbedded sandstone or sandy units occur. Williams (1943) named the upper Mississippian with a basal calcareous sandstone and thick middle and upper limestone sequence at Dry Lake section of Wellsville Mountain Brazer Limestone, dividing the Brazer Limestone into 5 units (Williams and Yolton 1945). Sando, Dutro and Gere (1959), in reevaluating the Brazer Limestone, found it to be mostly dolomite with a few beds of quartz sandstone and limestone near the top. Meager fossils give the lower two-thirds of the Brazer Dolomite an early Mississippian age with the upper third presumed to be upper Mississippian. It was then recommended by these authors that the Brazer Dolomite be restricted as a local term for Mississippian dolomite in the Crawford Mountains and that new names be applied to all upper Mississippian formations previously named "Brazer" Limestone. The lower two-thirds of the Brazer Dolomite has a fauna which is time equivalent to the Mission Canyon Limestone of the Madison Group.

It is of interest that work done at Wellsville Mountain by Sadlick (1955) correlated units 1-5 of Brazer Limestone with upper Mississippian stratigraphy in central Utah: unit 1, Humbug Formation; unit 2, lower limestone; unit 3, Long Trail Shale; unit 4, upper limestone of the Great Blue Formation; and unit 5, Manning Canyon Shale. Beus (1958) and Gelnett (1958) then mapped out the geology present at Wellsville Mountain but used Brazer Limestone to include the upper Mississippian. Williams (1958), in the *Geologic Atlas of Utah, Cache County*, believed terminology of the central Utah Oquirrh Range appeared to be applicable at Dry Lake section, that units 1-5 of Brazer Limestone seemingly were correlative as follows: unit 1, Humbug Formation; unit 2, lower Great Blue; unit 3, the equivalent of Long Trail Shale; unit 4, upper Great Blue Limestone; and unit 5, largely equivalent to Manning Canyon Shale. Carr and Trimble (1961) then correlated Great Blue Formation into the Deep Creek Mountains in southern Idaho from the southern Oquirrh Range in central Utah and used the name *Great Blue Formation* to describe parts of the "Brazer" Limestone at Dry Lake section of Wellsville Mountain. However, Williams (1963) in a review of Carboniferous and Permian rocks of Utah still cited the upper Mississippian at Wellsville Mountain as "Brazer" Limestone, but noted Great Blue Formation to be present farther north at Blue Springs Hills and slightly farther west at Little Mountain. Also, Beus (1963, 1968), in studies at Blue Springs Hills and Samaria Mountain, correlated Humbug Formation and Great Blue Formation from Promontory Range, Utah, through Dry Lake section of Wellsville Mountain into the northern Utah and southern Idaho study areas. Beus also used the term "Brazer" Limestone to describe the upper Mississippian at Dry Lake section but stated units 1-5 were correlative with Humbug Formation, Great Blue Formation, and Manning Canyon Shale. Beus (1975 pers. comm.) now believes Great Blue Formation to be the correct terminology for certain units of the "Brazer" Limestone at Wellsville Mountain. Previously, Olson (1956, 1960) had correlated Madison Limestone (now Gardison), Deseret Limestone, Humbug Formation, Great Blue Formation, and Manning Canyon Shale from central Utah to outcroppings at Promontory Range, Utah. Olson (1960) then tentatively correlated Humbug Formation, Great Blue Formation, and Manning Canyon Shale as equivalents to various units of the Brazer Limestone. Foutz (1966) also correlated Humbug Formation, Great Blue Formation, and Manning Canyon Shale from central Utah to Dry Lake section. Butkus (1975) then applied the names Humbug Formation and Great Blue Formation to upper Mississippian rocks at Dry Lake section, which he correlated with the same rock types in central Utah. Whether informally or formally printed, terminology at Wellsville Mountain for Mississippian rock sequences should be as follows upward through the section: Lodgepole Limestone, with possible upper portion Deseret Limestone; unconformity; Humbug Formation; Great Blue Formation; with Manning Canyon Shale present at Dry Lake section whereas to the north at Deweyville section an unconformity separates Great Blue Formation from overlying Oquirrh Formation. Relationship of this upppermost Mississippian unconformity from Deweyville section to Dry Lake section farther south is vague, but apparently lies above the Manning Canyon Shale.

Stratigraphy of the measured Deweyville section at Wellsville Mountain began at the base of Humbug Formation upward through Great Blue Formation into basal Oquirrh Formation. Base of the Humbug Formation is an unconformity separating it from the upper units of either Lodgepole Limestone or Deseret Limestone. Humbug Formation was named by Tower and Smith (1898) in the Tintic Mining District from the Humbug mine to what Spurr (1895) referred to as the Lower Intercalated series below Great Blue Limestone in the Mercur Mining district. Humbug Formation at Wellsville Mountain is largely a covered slope former making detailed study difficult; 24 samples were collected from Humbug Formation at slightly more resistant units which formed low outcrops or ledges. General field color of Humbug Formation is grayish orange for weathered surfaces and medium through dark medium gray with some light olive gray on fresh broken surfaces. Rock type in field description is calcareous sandstone with dolomite more abundant locally than calcite. Clastic to carbonate ratio of Humbug Formation in most collection sites was found to be nearly 1:1 with occasional increases in either admixture changing the name applied to that particular portion of the stratigraphic column (fig. 3). Humbug Formation was divided into six field units with changing composition within individual units occasionally moving across the clastic or carbonate fence to a different rock name. Sedimentary structures are rare with only some stratification observed, most samples being homogeneous. Uppermost contact of Humbug Formation with Great Blue Formation is placed at the abrupt change from a calcareous dolomitic sandstone to a medium to dark gray, sandy limestone with some bedded chert, 149.4 m into the measured section. The contact is distinct in the field with a sudden steepening of the gentle slope, the exact contact being covered by a 6.1-meter transitional area. The contact between the Humbug Formation and Great Blue Formation reflects an abrupt dominance of carbonates with clastics of lowered importance, with progressively thinner interbeds of Humbug Formation-like units upward within lower portions of Great Blue Formation.

Great Blue Limestone was first named by Spurr (1895) in the Mercur Mining district with a Lower and Upper Intercalated series bounding rock sequences "which may be called Great Blue limestone," with the shale sequence within Great Blue Limestone formally named Long Trail Shale by Gilluly (1932) from Long Trail Gulch in Ophir Canyon. Limestone units above and below Long Trail Shale were left as informal lower limestone member and upper limestone member. G. H. Girty (in Gilluly 1932) established an upper Mississippian age for the Great Blue Limestone. Great Blue Formation was given formal status by Morris and Lovering (1961). In a study of the stratigraphy of the East Tintic Mountains they formally named four members which are in ascending order: the Topliff Limestone member, the Paymaster member, the Chiulos member, and the Poker Knoll Limestone member from

Topliff Hill and adjoining knolls. Long Trail Shale was found by Morris and Lovering (1961) not to be equivalent to the Chiulos member, possibly being present only as one or more black shale beds in the lower portion of the Paymaster member. Crittenden (1959) has also pointed to possible confusion which may result by attaching Long Trail Shale to any shale sequence found near the middle of the Great Blue Formation.

At Deweyville section of Wellsville Mountain, Great Blue Formation grades out of the slope-forming Humbug Formation into an interfingering sandy limestone, limestone wackestone, and calcareous dolomitic sandstone sequence to form a steep ledge and slope former, medium to thick bedded. Clastic and dolomite admixtures interfingering with limestone sequences disappear upward in the section as the first prominent cliff-forming sequence is reached 116.5 meters above the base of the formation. Wackestone and sandy limestone are medium through dark gray in color, in comparison to lighter calcareous dolomitic sandstone intervals. A small but laterally extensive biostrome accumulation of colonial corals forms the base of the first prominent cliff sequence. These corals appear to be similar to what Parks (1951) classified as *Lithostrotion whitneyi* Meek in the lower portion of Great Blue Formation at Dry Lake section. Since that study, more detailed regional coral zonations and classifications have been put forth but are not discussed herein. Suffice it to say that the coral-bearing unit permits correlation with units at Dry Lake section.

The lower part of Great Blue Formation forming this first prominent cliff sequence is a series of three small cliffs, separated by small, steep ledge areas, which stand out boldly from the mountain front. Rock types in the cliff-forming sequence are limestones of wackestone and packstone, composed of skeletal rock-building constituents with some pellets and a few thin beds of oolites. Color of the rocks is mostly dark gray on fresh surfaces with medium through medium dark gray on weathered surfaces. Bedding is variable from occasional thin- through very thick-bedded sequences, with individual units abruptly changing lithology vertically (fig. 3). Thinner coral-bearing beds are developed and scattered through parts of the section, all being groupings of solitary rugose corals. Weathering in places below some, but not all, of these solitary coral horizons has produced concave weathered features within the mountainside. Below these few developed concave features, the rock is weaker and breaks into large chips. Above the first cliff sequence a broad, covered slope former separates the first cliff sequence from a second well-developed cliff sequence. In the slope former, wackestone is the common rock type, with two thin oolite beds present, and then grades out of a slope sequence into a ledge and slope former leading up to the second well-developed cliff former. Skeletal and pelletal constituents are the rock builders. At the base of the second cliff-forming sequence is found the most well-developed rugose solitary-coral concentration in the section, all corals resting horizontally in a bed that is 1-2 meters thick. This coral-bearing unit grades upward into oolitic limestone, which makes up most of the second cliff-forming sequence.

Oolitic limestone forms the steep second cliff former and is the most resistant unit in the measured section. A few thin coral-bearing beds are interbedded with oolitic limestone. Upon grading out of oolitic limestone the steep cliff abruptly gives way to a ledge and slope former. Oolitic limestone occurs 375.5 meters above the base of the measured section and 110 meters up from the base of the first cliff sequence and is 24.4 meters thick. Within the ledge- and slope-forming sequence above the second cliff former, limestone varies between mudstone and wackestone with changing amounts of skeletal and some pelletal constituents. At the top of this ledge- and slope-forming sequence is bedded chert; bedding throughout medium and thick, with medium-dark-to-dark-gray color common to the rocks, the weathered surface being lighter gray. The total thickness for this sequence is 21.6 meters, with the upper contact forming an abrupt break in the section and interbedded claystone and limestone units above the break. This upper contact marks the top of the lower limestone member that Gilluly (1932) described in central Utah as lower limestone member and that Sadlick (1955) described at Dry Lake as Lower Limestone. Recent work by Butkus (1975) refers to this lower limestone sequence as the Topliff limestone member, after Morris and Lovering (1961). This lower limestone sequence has a total thickness of 272.2 meters.

A slope former is next composed of interbedded claystone and limestone units. Claystone is thin bedded and very weak, with most units covered by 0.5 meter of slope cover and with two calcareous shale units at the top. Limestone is present as small lenses within claystone units, forming micrite nodules the size of a fist. Limestone units interbedded between claystone units are mudstone, wackestone, and packstone composed of skeletal and some pelletal rock-building constituents, and are medium to thick bedded. This sequence is terminated upward at the contact of the highest calcareous shale unit, with the ledge- and slope-forming sequence now becoming a more continuous, very steep slope former of limestone only. The interbedded claystone and limestone sequence is 61.6 meters thick. Claystone is pale to dark yellowish brown to light olive gray in color, with exposed limestone units mostly dark gray. This interbedded claystone and limestone sequence is separated from both the lower and upper limestone sequences and was formally named Long Trail Shale by Gilluly (1932) and proposed as such at Wellsville Mountain by Sadlick (1955). Later work by Crittenden (1959) and Morris and Lovering (1961) have put restraints on lateral use of the term *Long Trail Shale,* and Butkus (1975) referred to this sequence as median shale member; I favor the latter usage.

Upper limestone sequences above interbedded claystone and limestone units form a rather steep continuous slope with a transition approximately halfway where steep ledge- and slope-forming topography to near-continuous ledge sequence leads up to the uncomformable top of the Great Blue Formation The lower half forms units of limestone only with local chert beds becoming more common; bedding is medium to thick, with some thin. Rock type is mostly wackestone, with small amounts of mudstone and packstone. Color is medium through dark gray for both weathered and fresh surfaces. When the steep ledge and slope-forming sequence is reached, halfway through the upper limestone sequence, limestone units give way to clastic interbeds of sandy limestone and sandstone which weather to yellowish-brown tones and are traceable across the ridge to adjacent ridges. Upward into this upper part of the sequence calcareous and sandy dolomite also becomes interbedded with limestone, sandy limestone, and sandstone lithologies. Individual units become thin bedded in parts of the upper half of this sequence with much thicker individual units being medium to thick bedded. Certain units contain chert concentrations that comprise as much as 40 percent of the rock. Coral beds in this upper limestone sequence are very scarce in comparison with the lower limestone sequence. Total thickness of the upper limestone sequence is 206.6 meters, and it can be divided into lower and upper halves. The lower half is completely limestone 84.4 meters thick, and the upper half is interbedded limestone, sandy limestone, calcareous to sandy dolomite, and sandstone and is 122.2 meters thick. Gilluly (1932) called this informally the upper limestone member, and Sadlick (1955) referred to it as Upper Limestone. Butkus (1975) uses the more acceptable terminology of upper limestone member.

At Deweyville section of Wellsville Mountain, there are three recognizable but informal members composing the Great Blue Formation. (1) Interbedded claystone and limestone sequences at Wellsville Mountain originally referred to as Long Trail Shale by Gilluly (1932) are herein referred to as a medial shale member as described by Butkus (1975). (2) Upper limestone sequences with interbedded lithologies in the upper portion retains herein Butkus's (1975) designation upper limestone member. (3) Lower limestone sequences were originally named lower limestone member by Gilluly (1932) for limestone sequences below Long Trail Shale. Sadlick (1955) then used the term *Lower Limestone* at Wellsville Mountain for this sequence. Later, Morris and Lovering (1961) formally named lower limestone sequences in the East Tintic Mountains the Topliff limestone member, which name Butkus (1975) applied to all lower limestone sequences within the Great Blue Formation. I believe simpler terminology for the lower limestone sequence at Wellsville Mountain is preferable and will use the term lower limestone member, as described by Gilluly (1932).

PETROLOGY AND PETROGRAPHY

Seven distinct facies are recognized to characterize Great Blue Formation at Deweyville section of Wellsville Mountain (fig. 4). These facies do not crosscut the three informal members but subdivide them into lithologies, each of which characterizes one facies only. Both petrologic and petrographic description of Great Blue Formation will be discussed on a facies-to-facies basis.

Lower limestone member is divided into four facies, which are, in ascending order: (1) wackestone dolomite sandstone facies; (2) wackestone-packstone facies; (3) oolite facies; and (4) mudstone-packstone facies. Median shale member is herein regarded as the fifth facies, an alternating sequence of claystone and limestone subfacies, forming a slope and grading into a ledge and slope former near its upper terminus; it is named the mudstone-wackestone claystone facies. Upper limestone member is divided into two facies: (1) wackestone facies, and (2) mudstone-packstone dolomite sandstone facies. Within all these, to a greater or lesser extent, smaller, occasionally minor, subfacies occur and interfinger with the dominant subfacies. Subfacies are of minor significance individually, but collectively are significant.

Wackestone Dolomite Sandstone Facies

At the base of the measured section of Great Blue Formation, the more characteristic lithologies present in Humbug Formation, such as calcareous dolomitic sandstone, calcareous sandy dolomite, and, to some extent, sandy limestone interfinger with limestone sequences of mudstone and wackestone which are more typical of Great Blue Formation. Though a sudden increase in limestone marks the base of Great Blue Formation, lithologies similar to Humbug Formation interfinger within parts of the wackestone dolomite sandstone facies. Three subfacies are present in this lowermost basal facies; (1) limestone and sandy limestone subfacies, (2) sandstone subfacies, and (3) dolomite subfacies. These subfacies experience oc-

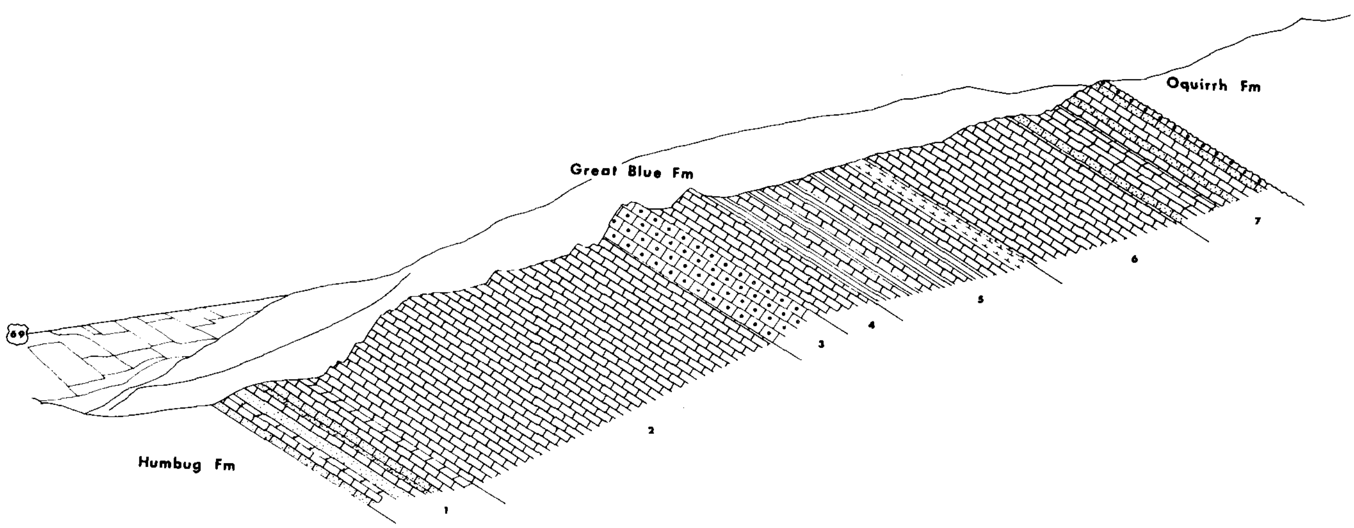


FIGURE 4.—Ridge outline and seven facies composing Great Blue Formation just north of the measured Deweyville section, north side of Holdaway Canyon. (1) Wackestone dolomite sand stone facies, (2) wackestone-packstone facies, (3) oolite facies, (4) mudstone-packstone facies, (5) mudstone-packstone claystone facies, (6) wackestone facies, and (7) mudstone-packstone dolomite sandstone faciees.

casional transitional shifts across a compositional fence to change rock names; for example, a calcareous, dolomitic sandstone becomes less sandy and more dolomitic to form a calcareous, sandy dolomite.

The first basal unit is a sandy limestone and wackestone 26.8 meters thick, with occasional chert bands (fig. 5). A sample was taken from just above the base of the contact between Humbug Formation and Great Blue Formation and, though essentially a limestone, it contains clastics of fine-grained-to-coarse silt size to form a sandy mudstone. Matrix is formed of micrite mud recrystallized to a fine microspar, but is not visible at low magnification. Only one shell of skeletal material is visible in the center of figure 5, and it has been altered by silica replacement. Clastics are both equant and elongate and well sorted, and nearly all are quartz grains with minor authigenic feldspar. Very small amounts of fine detrital crinoid columnals are scattered through the sample. A second thick unit overlies the first with a composition which varies from calcareous, highly sandy dolomite to a calcareous, dolomitic sandstone with only small increases in calcareous admixtures to form limestone. This second thick unit is similar to the underlying Humbug Formation and probably represents a minor shifting back to conditions of clastic sediment formation and deposition. Above these two thick units the remainder of this facies is formed of thinner units 1.2-4.6 meters thick, which interchange lithologies frequently and form a continuous ledge and slope-forming sequence. Sandstone subfacies is generally fine to very fine grained to coarsely silty and interfingers with the facies at different elevations (fig. 6). In the figure quartz grains are essentially grain supported with grain-to-grain boundaries impinging into each other to form welded contacts. Matrix is present as both calcite and dolomite. Calcite is more common and impinges into clastic grains to leave etched or serrated grain edges. Dolorhombs have formed, with darker (stained) spots of calcite within the rhombs of dolomite showing the latter to have undergone a small amount of dedolomitization (just right of D in fig. 6). Dolomite subfacies is always sandy, with calcite only slightly common to rare. Iron oxide is common in 1-3 percent amounts when dolomite is present (fig. 7). Individual dolorhombs have been diagenetically enlarged and impinge into clastic grains, with each rhomb a brown color because of iron as an impurity in the dolomite structure. In rare instances larger dolorhombs show development of secondary dolomite overgrowths leaving a smaller rhomb within a larger one, suggesting two periods of dolomitization (arrow in fig. 7). Outer overgrowths are slightly more clear and less tinged with iron, suggesting that the second period of dolomitization was accompanied by less iron. Most dolorhombs are well developed and have very sharp boundaries which impinge without alteration of the rhomb shape directly into clastic grains. Dolomite development is therefore interpreted as secondary because of diagenetic dolomite development. Occasionally dolomite will form only part of the lithology present, being confined to the lower part of a particular unit composed of limestone upsection.

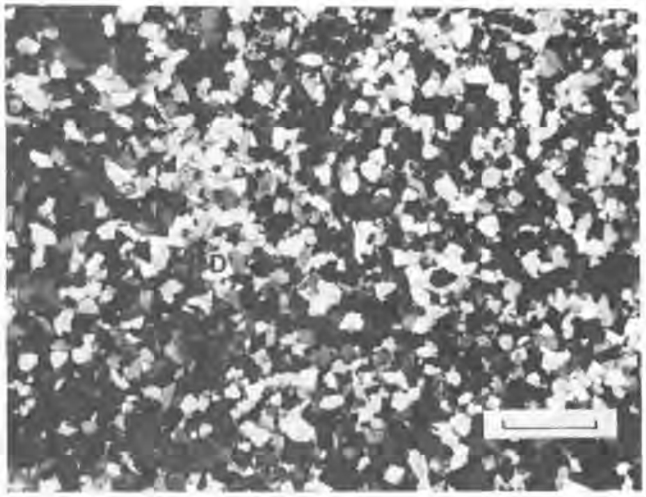


FIGURE 6.—Photomicrograph (crossed nicols) of dolomitic, calcareous sandstone, totally stained with Alizarin Red S. Scale is 1 mm.

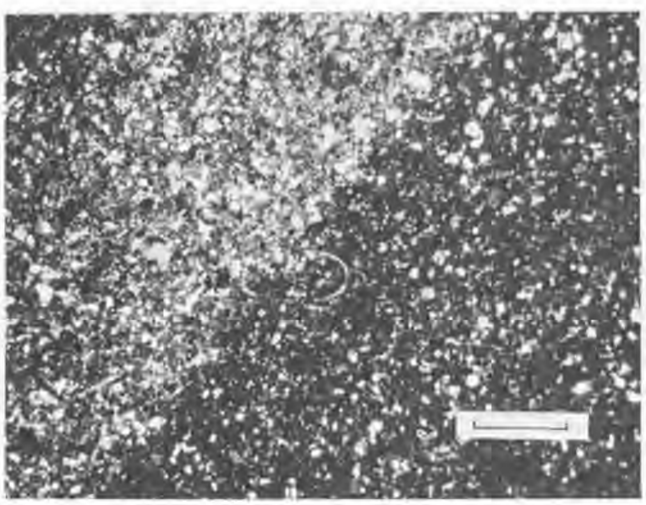


FIGURE 5.—Photomicrograph (crossed nicols) of sandy limestone subfacies, right half stained with Alizarin Red S. Scale is 1 mm.

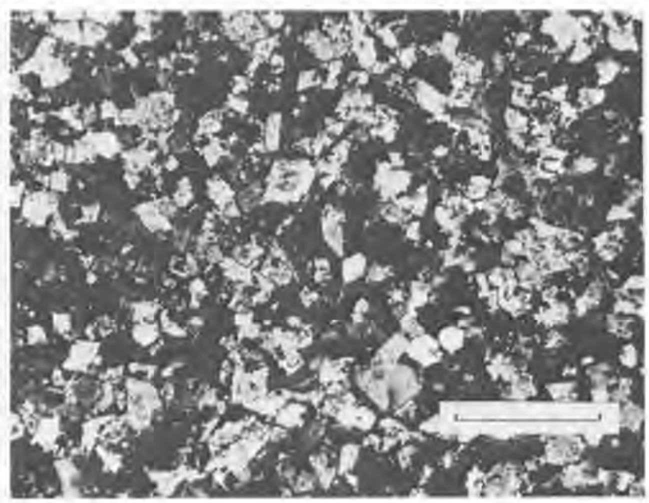


FIGURE 7.—Photomicrograph (crossed nicols) of dolomite. Arrow pointing to dolorhomb which has undergone two periods of development. Scale is 0.25 mm.

Wackestone-Packstone Facies

The next facies is characterized by wackestone and packstone, with dolomite and clastic constituents absent. Wackestones and packstones are composed of skeletal and pelletal rock-building constituents set in fine-grained micrite mud. These two limestone types form a dominant subfacies accompanied in the section by a second subfacies composed of one colonial coral biostrome and several minor, solitary coral-bearing beds. A third subfacies is also present in the form of three small oolite beds, each 0.3 meters thick. Upsection these three subfacies interfinger in an orderly manner to produce the wackestone-packstone facies. The lower part of this facies serves as an example of this orderly arrangement. At its base a sudden abundance of skeletal packstone marks the beginning of the facies, which rests upon a fine-grained calcareous, sandy dolomite of the wackestone dolomite sandstone facies (fig. 8). The contact forms the first unit within the wackestone-packstone facies and reveals skeletal packstone to give way to pelletal and detrital skeletal constituents of crinoid columnals with whole foraminifera to form a wackestone (unit 15, fig. 3). The second unit, 16, has at its base similar pelletal, less detrital, and more whole skeletal constituents of crinoid columnals and whole foraminifera, forming another wackestone which comprises the base of a 1.2-meter biostrome of colonial coral *Lithostrotion whitneyi* Meek (Parks 1951) (fig. 9). Therefore, in vertical sequence rock-building constituents are (1) detrital crinoid columnals with some pellets; then (2) pellets, detrital crinoids, and complete foraminifera; and finally (3) fewer pellets with larger crinoid detritus and whole foraminifera overlain by a colonial coral biostrome. Above the colonial coral biostrome a reverse sequence is present with complete crinoid columnals with some pellets and detrital algal micrite (unit 17, figs. 3 and 10). Here recrystallized crinoid columnals and echinoid plates, with only rare brachiopod shells, compose the skeletal constituents, with syntaxial rim cement development and formation of calcite twinning through both skeletal constituents and rim cement. Darker, irregularly shaped, oval areas consist of matrix composed of clotted algal micrite. Grayish areas

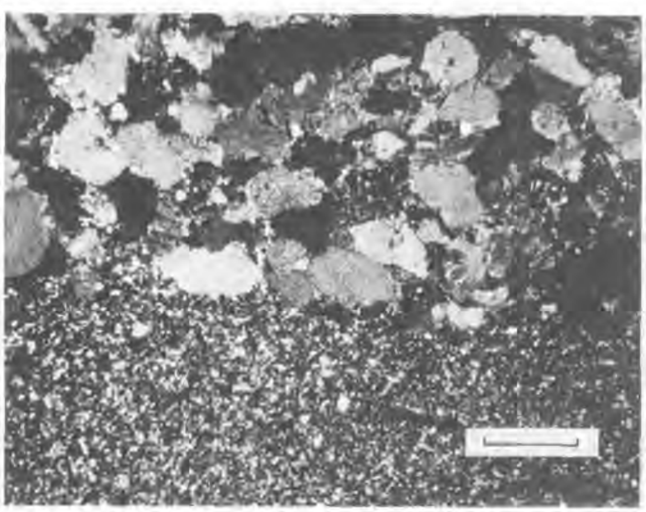


FIGURE 8.—Photomicrograph (crossed nicols) of contact between lower wackestone dolomite sandstone facies and overlying wackestone-packstone facies. Scale is 1 mm.

FIGURE 9.—Coral biostrome-forming base of the first prominent cliff-forming sequence. Staff is marked in 1-foot intervals.

are more algal deficient with development of microspar beginning to invade the matrix. Skeletal constituents in the photomicrograph (fig. 10) reflect a less detrital, more complete unbroken appearance.

The next unit, 18, 0.6 meter thick, is a pelletal-rich wackestone with no complete crinoid columnals but some crinoid detritus (fig. 11). Here skeletal constituents are much more detrital than in the previously described sample, and most, though recrystallized, are crinoid columnals and brachiopod or pelecypod shells. Darker areas are algal pellets, very abundant in this photomicrograph. Both skeletal and pelletal constituents are set in a matrix of smaller skeletal detritus and micrite mud which in algal-poor areas are recrystallized to microspar but are not evident at low magnification. These subfacies interfinger back and forth, establishing different rock units in an orderly pattern.

Slightly higher in the section, unit 26, another subfacies within the same facies, consists of pelletal and highly recrystallized crinoid columnals and echinoid plates in its base. Upsection this unit grades upward through two very small solitary coral-bearing beds and then into the

FIGURE 10.—Photomicrograph (crossed nicols) of clotted micrite, skeletal packstone. Scale is 1 mm.

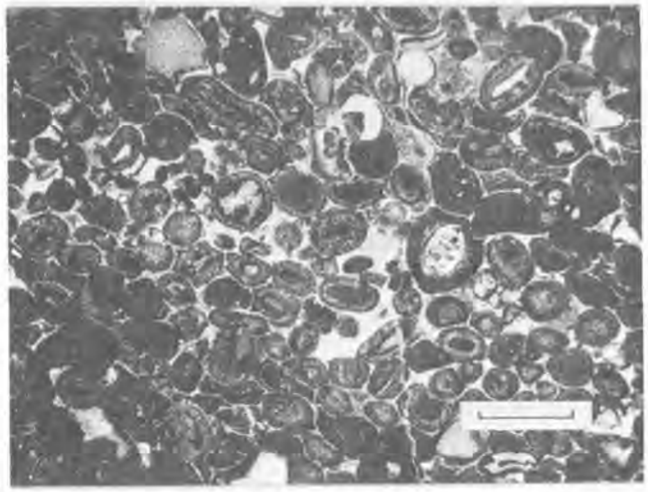


FIGURE 12.—Photomicrograph (crossed nicols) of oolitic limestone. Ooids are crushed and impinged into. Scale is 1 mm.

first thin oolite bed in the measured section (fig. 12). These are mostly true oolites, containing nuclei in their centers around which several algal envelopes developed. The ooid just right of center (fig. 12) appears to have an inner and an outer coating which could represent two periods of its development. Most ooids reflect an outward fibrous appearance associated with algal envelopes, and in many two sequences are apparent. Nuclei forming the center of most of them are detrital skeletal pieces of both bryozoan and crinoid columnals. All dark noncoated bodies in the photomicrograph are detrital, rolled algal micrite. An interesting feature to be observed in this photomicrograph (fig. 12) is the crushing or distortion which all ooids and rolled micrite have undergone. Crushing did not fracture individual ooids but only deformed them. This observation is interpreted as indicating an early event that took place before lithification. If this subfacies is combined with the subfacies described previously, subfacies changes may occur as follows: (1) pellets possibly with detrital skeletal material; (2) less pelletal and more skeletal constituents that are less detrital; (3) thin coral-bearing beds with rare to common amounts of corals; and (4) thin oolite beds. Mudstone, occasionally interfingering within skeletal and pelletal units, is composed of fine-grained micrite mud with a few whole brachiopod shells (fig. 13). A trace fossil grazing path cut through the mud, leaving behind a trail filled with both micrite mud and microspar and initiating development of very small calcspars of granular sparry calcite. The remainder of the sample is composed completely of one of the most fine-grained micritic mudstones to be found in the measured section.

Throughout this facies the three subfacies interfinger vertico-laterally several times; oolite beds occur only three times, with thin coral-bearing beds being developed more often. The most common lithology is the shifting rock-building constituents of skeletal and pelletal material

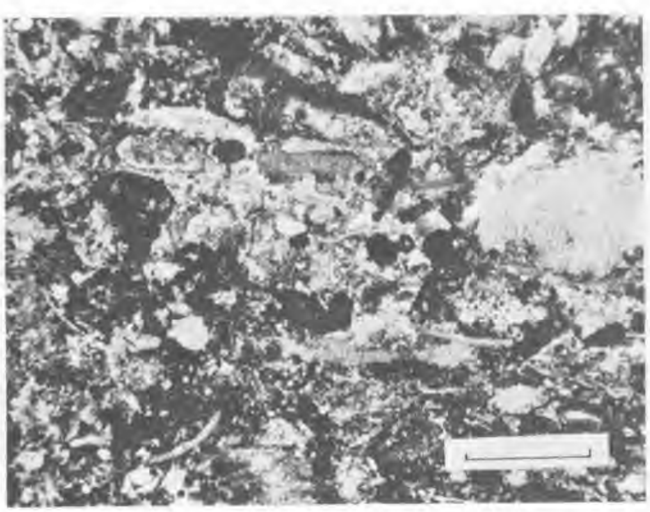


FIGURE 11.—Photomicrograph (crossed nicols) of skeletal, pelletal wackestone. Scale is 0.5 mm.

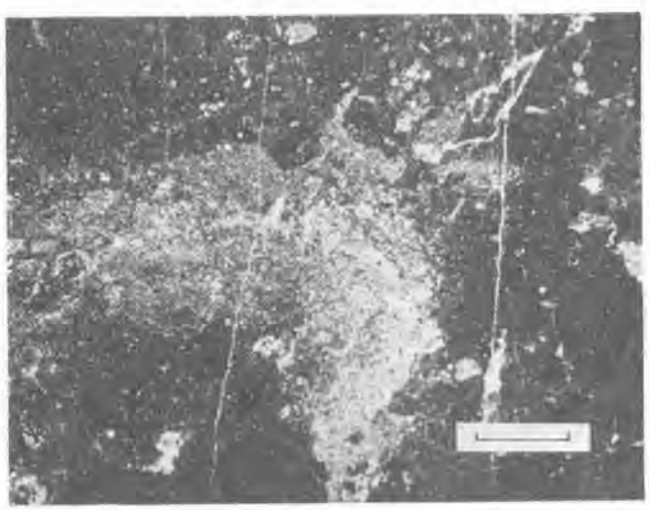


FIGURE 13.—Photomicrograph (crossed nicols) of mudstone with a trace fossil burrow path cutting through the photo. Scale is 1 mm.

which form the dominant wackestone and packstone subfacies. It is not intended herein to account for each change that occurred, but rather to show that during deposition there was continuous interfingering back and forth of these subfacies to produce the larger, more recognizable wackestone-packstone facies.

Oolite Facies

The last rock unit of the wackestone-packstone facies is a pelletal-skeletal limestone which grades upward into well-developed solitary coral-bearing beds within the unit, 375.5 meters above the base of the measured section, forming the base of the second and most well-developed cliff. Stratigraphically above these coral-bearing beds, ooids become part of the rock-building constituents and define the base of the oolite facies (fig. 14). In the figure the rock composition consists of both oolites and skeletal detritus, but individual ooids are not the dominant rock builders. Dominant amounts of skeletal detritus are composed of small pieces of bryozoans, echinoids, and brachiopod shells, still recognizable, with finer nondescript detritus of fragmental skeletal constituents comprising most of the matrix. The ooid (fig. 14) above center has a nucleus composed of an entire foraminifera, encased in several algal envelopes. This change from a near-solitary coral colony upward into an oolite sequence 24.4 meters thick appears to substantiate earlier observations of a possible relationship between oolite development and coral formation. Within the oolite facies a few small, thin coral-bearing beds are present, reflecting a small amount of interfingering of oolite facies with coralline limestone. Within ooid-rich portions of this facies there occur both true oolites (most with nuclei coated by algal micrite) and intraclastic skeletal coated grains, both set in original pore cement of granular sparry calcite with some slightly recrystallized algal micrite mud (fig. 15). Intraclastic skeletal constituents forming ooid and coated-grain nuclei contain more delicately preserved bryozoan detritus because bryzoans are easily broken up by

FIGURE 15.—Photomicrograph (crossed nicols) of oolitic limestone with skeletal debris forming nuclei centers and well-developed pore cement. Scale is 1 mm.

wave or current action if not encased within other material. What is evident in figure 15 is the excellent development of the primary pore cement of interlocking sparry calcite which forms most of the matrix. The presence of small amounts of micrite mud (forming dark, cloudy areas between ooids and coated grains) reflects that, while primary pore cement was being precipitated, the environment must not have been totally receptive to its development because of accompanying micrite mud. The oolite facies is not always completely composed of ooids. It sometimes consists of both ooids and detrital skeletal constituents. Or it may be made up of detrital skeletal material either as coated grains or as intraclasts where a slight interfingering occurs to form no coating on detrital skeletal constituents (fig. 16). Here oolites and coated grains rest in a matrix of detrital skeletal intraclasts and slightly recrystallized micrite mud. Interfingering of an oolite environment with a skeletal environment appears to be present in this one sample, where some skeletal constituents form nuclei

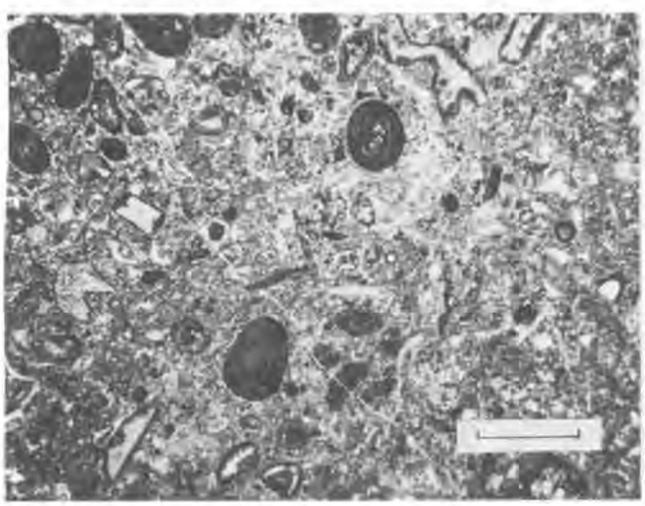


FIGURE 14.—Photomicrograph (crossed nicols) of oolitic limestone, revealing ooids and detrital skeletal constitutents to be rock builders. Scale is 1 mm.

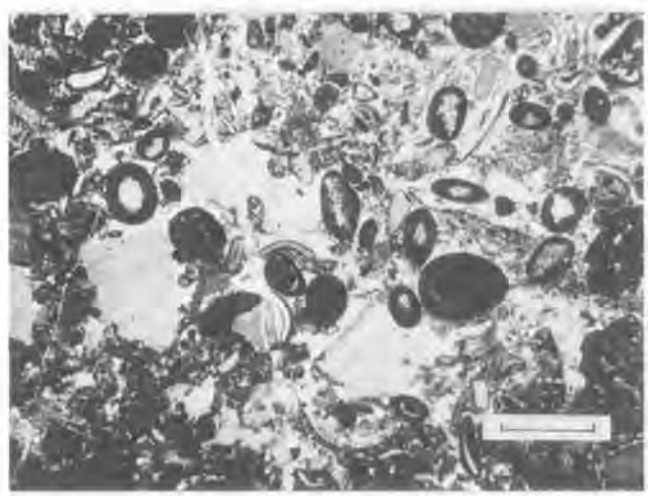


FIGURE 16.—Photomicrograph (crossed nicols) of oolitic limestone, revealing ooids and skeletal detritus. Some skeletal debris are without algal envelope development. Scale is 1 mm.

for ooids and coated grains whereas others are detrital with no algal envelope coatings. Thus within the oolite facies, though only slightly apparent in the field, most of the facies was deposited, not in an oolite-generating zone, but in an interfingering area where both ooids and detrital skeletal constituents were produced and accumulated.

Mudstone-Packstone Facies

At the top of the oolite facies the cliff former abruptly gives way to a ledge- and slope-forming sequence which varies in lithology from mudstone to wackestone and packstone, all composed of varying amounts of skeletal and pelletal rock-building constituents set in fine-grained micrite mud. In this facies there are two subfacies: (1) a fine-grained, slightly skeletal mudstone; and (2) a skeletal and pelletal wackestone to packstone. It is of interest that, as the oolite facies gives way upward to this mudstone-packstone facies, a small, thin, solitary, coral-bearing bed has developed where oolites gradually diminish in importance and grade upward into rocks of this facies. Within the mudstone-packstone facies there is occasional development of small, thin, solitary, coral-bearing beds that form within wackestone and packstone subfacies. Coral-bearing beds are considered to be a rarely formed third subfacies which interfingers only within the wackestone and packstone subfacies (see fig. 3).

Mudstone subfacies is composed of very fine-grained micrite mud. Its high resistance to recrystallization suggests that it is of an algal origin. Mudstone contains only small amounts of skeletal detritus and occasionally whole shells of pelecypods (fig. 17). At the left of this figure one complete pelecypod shell is preserved with open space filled by sparry calcite. Calcspars fill shell interstices, and the shell itself is totally recrystallized. Other shell halves are scattered within the sample and are also recrystallized, with small detrital pieces of skeletal material resting in fine micrite mud. A fracture traversing the sample has intersected a shell half which is totally recrystallized with fibrous sparry calcite around shell edges. Irregular-to-circular patches are areas recrystallized by neomorphism with development of various sizes of interlocking granular sparry calcite. On rare occasions sponge spicules of Hexactinellide sponges (Rigby, pers. comm., 1976) are scattered within the micrite mud. Individual spicules are replaced by sparry calcite but are not apparent at this magnification.

FIGURE 17.—Photomicrograph (crossed nicols) of mudstone. Scale is 1 mm.

Wackestone and packstone subfacies are formed from detrital skeletal and pelletal rock-building constituents (fig. 18). Most dark areas of this figure are algal pellets which are the dominant rock-building constituent in this sample. Lighter areas are detrital skeletal material: crinoid columnals, echinoid plates, and some shell debris, all recrystallized with syntaxial rim cement overgrowth development which has invaded algal-poor areas of matrix and recrystallized them together with skeletal material. As noted previously, when high amounts of pellet constituents are present, most skeletal material is more detrital and less common. Pellets are not incorporated or recrystallized by syntaxial rim cement overgrowth but have left irregular edges associated with most detrital skeletal constituents because of a reverse impingement outward of rim cement development. Where wackestone and packstone subfacies begin to show greater accumulations of more diverse skeletal constituents, pelletal constituents become less important (fig. 19). Although still detrital, skeletal constituents are notably larger and less fragmental than other previously described thin sections of this subfacies. In figure 19 an oval echinoid plate reveals some of its primary structure though neomorphism has affected it and forms a recrystallized matrix surrounding it. Delicate bryozoan pieces are present, with dark areas of bryozoan structure filled with micrite mud whereas larger openings between actual parts of the skeletal structure reveal development of small-granular sparry calcite. Black oval-to-elongate-oval areas are detrital algal micrite rolled to their present shapes. Matrix reveals some dark areas of micrite mud surrounded by granular sparry calcite. Elsewhere lighter areas in the matrix are due to recrystallization of a former matrix, and possibly some original pore cement, into large patches of sparry calcite.

Mudstone-Wackestone Claystone Facies

Median shale member composes the complete mudstone-wackestone claystone facies with boundaries of the member also representing facies boundaries. Within this facies very fine clay minerals and clastics become prominent but appear different from clastics present in the wackestone dolomite sandstone basal facies. This facies may be subdivided into two subfacies: (1) limestone, which varies from mudstone to wackestone; and (2) claystone.

Claystone units, though most are buried 0.5 meter by slope cover and must be exposed by digging, reveal in thin section clay material and very small silt-sized clastics to be the rock builders, with clastics visible only at magnifications of 50X (fig. 20). Here fine laminations are visible with both clay minerals and fine clastics both too fine to be viewed at this low magnification. Small dark areas are patches of iron oxide which are finely disseminated throughout the sample. Occasionally clastics become slightly larger (in the coarse silt size) and may reflect flushing of clastics farther out into the depocenter. The last two clay-rich units are also enriched with calcium carbonate and have fissility developed to form calcareous shale units. Digging to buried claystone units also revealed the presence of micrite nodules the size of a fist as inter-

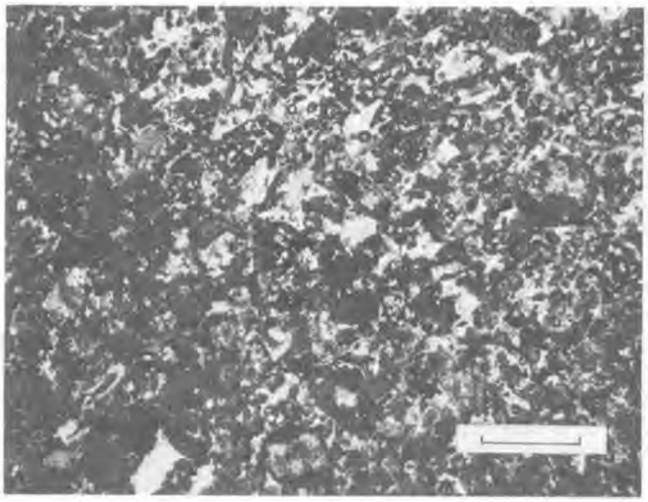


FIGURE 18.—Photomicrograph (crossed nicols) of skeletal, pelletal, wackestone to near packstone. Scale is 1 mm.

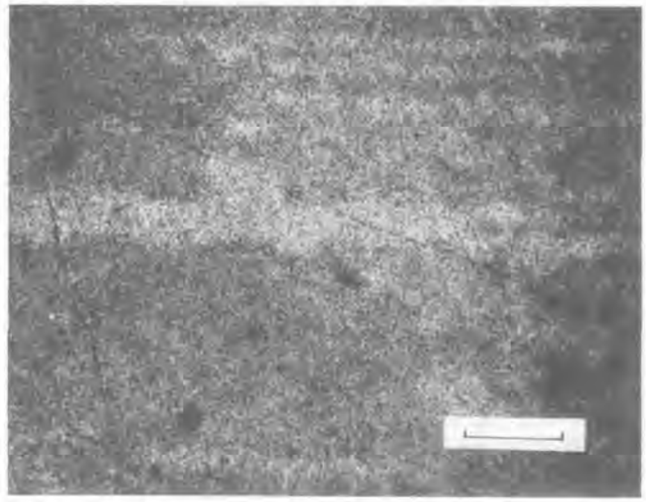


FIGURE 20.—Photomicrograph (plane light) of claystone. Note very fine laminations running through the photo. Scale is 1 mm.

beds within claystone units (fig. 21). Micrite nodules are formed of very fine-grained recrystallized interlocking microspar with pressure solution developed stylolites occasionally traversing the nodules. Along these stylolites there has been a secondary development of authigenic feldspar. Presumably the claystone units themselves supplied solutions during compaction and early lithification to form feldspars.

Limestone subfacies interfingers with both claystone and calcareous shale units to form either mudstone or wackestone. Constituents which form the two limestone rock types have various combinations of the following: (1) micrite mud; (2) detrital skeletal debris and micrite mud; (3) pelletal and detrital skeletal debris; and (4) dominantly detrital skeletal constituents. Through this mudstone-wackestone claystone facies, limestone subfacies change in both the degree of composition of rock builders and relative percentages of each. These varying limestone rock-building constituents probably follow the same orderly patterns established for facies in the lower limestone member if formed in the same basic environment, but within median shale member claystone interfingerings may serve to disturb the sequence enough only to suppose an orderly arrangement of micrite mud to pelletal limestone to more detrital skeletal-rich and pelletal-poor limestone.

Mudstone is mostly composed of dominant amounts of micrite mud, slightly recrystallized to form a very fine microspar, and small amounts of detrital skeletal debris, most of which is recrystallized with some silica replacement (fig. 22). In the figure larger detrital pieces are composed of recrystallized echinoid plates with impinging matrix surrounding each. Small sponge spicules, also recrystallized, form circular or spicule-shaped smaller detritus. Least common are irregular-shaped masses of bryozoan detritus which were strongly fragmented and have all been replaced by silica. Silica replacement of bryozoan material is a

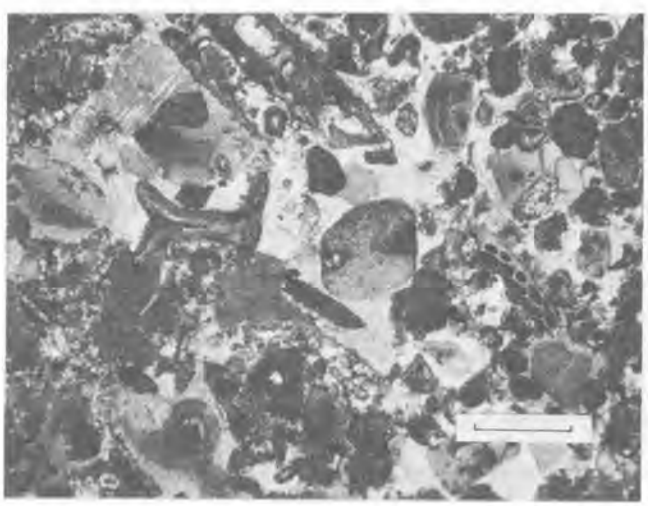


FIGURE 19.—Photomicrograph of pelletal, skeletal wackestone to near packstone. Scale is 1 mm.

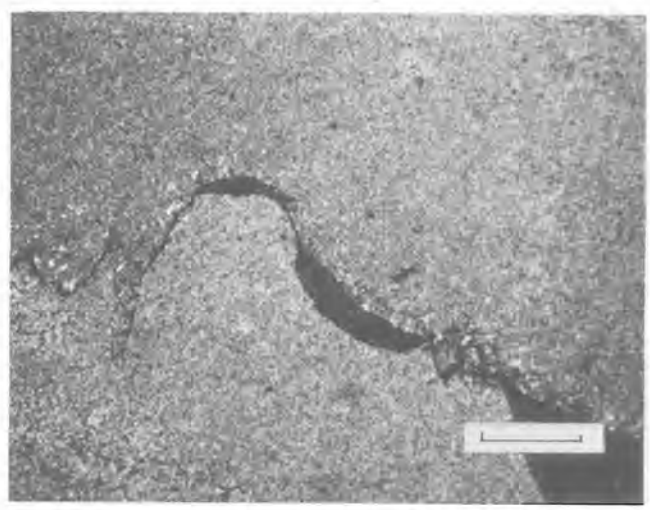


FIGURE 21.—Photomicrograph (crossed nicols) of a stylolite running through a micrite nodule with very small authigenic feldspars developed. Scale is 1 mm.

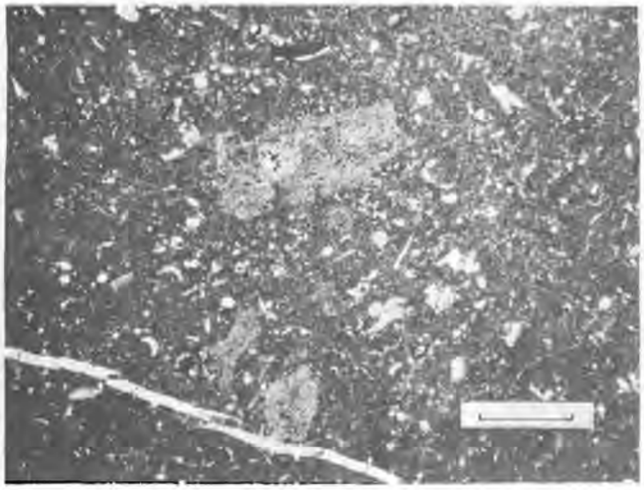


FIGURE 22.—Photomicrograph (crossed nicols) of mudstone. Scale is 1 mm.

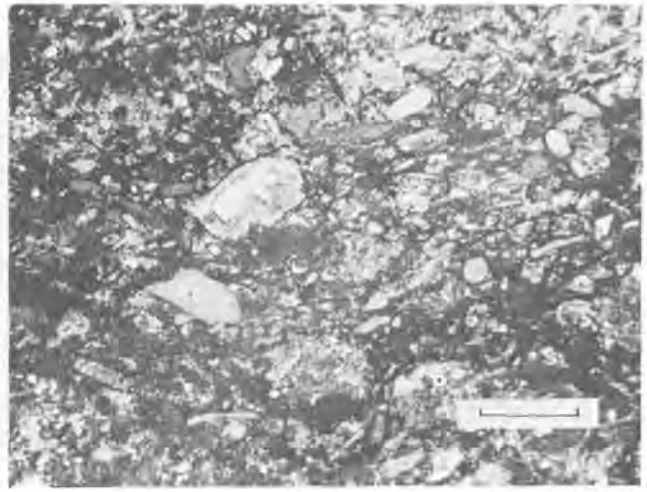


FIGURE 24.—Photomicrograph (crossed nicols) of skeletal wackestone with several small stylolites cutting through the photo. Scale is 1 mm.

feature noted in most thin sections. Bryozoan detritus apparently had a stronger tendency to be replaced by silica than did other detrital skeletal pieces. Certain portions of limestone sequences show pellets to be the dominant rock-building constituent with subordinate amounts of very fine skeletal detritus (fig. 23). Dark, small oval areas of this photomicrograph are algal pellets, and lighter areas are skeletal debris of brachiopod or pelecypod shells with smaller nondistinct skeletal debris intermixed with pellets. Neomorphism has recrystallized skeletal constituents with a few small interlocking calcspars formed locally within the sample. As micrite mud (mudstone) and pelletal rich (pelletal limestone) become less common, larger detrital skeletal constituents become the dominant rock-building constituents (fig. 24). Detrital skeleton-rich limestone sequences have varying amounts of crinoid columnals, echinoid plates, brachiopod or pelecypod shells, and whole foraminifera as the rock-building constituents. In figure 24 most of the detrital skeletal constituents have several

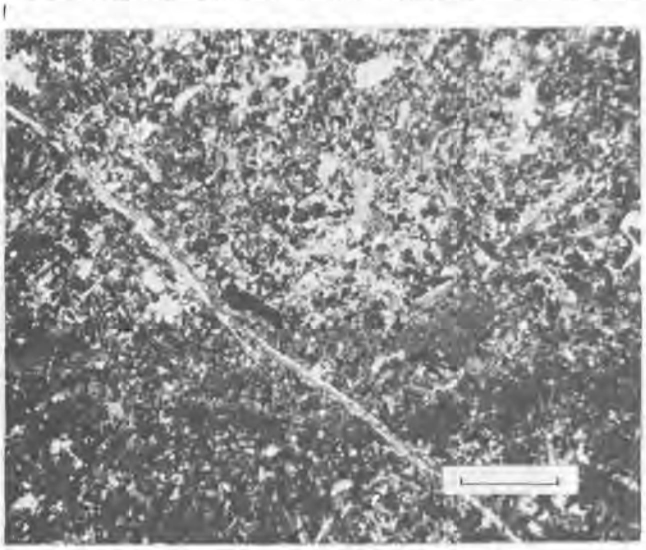


FIGURE 23.—Photomicrograph (crossed nicols) of skeletal, pelletal wackestone. Scale is 1 mm.

small stylolites cutting through the thin section, which have reshaped most skeletal boundaries highly. Matrix is composed of micrite mud, which has been partly recrystallized to form a very fine microspar or coarse textured micrite mud, with lighter portions of matrix where high amounts of silica have replaced small nondistinct skeletal detritus. Several stylolites cutting through areas of silica replacement could have served as pathways for replacing solutions which migrated through the sample and replaced more highly susceptible skeletal debris.

Wackestone Facies

Wackestone forms most of the lower portion of the upper limestone member. It is the most abundant rock type in this facies with less common amounts of mudstone and packstone. Rock-building constituents vary in amounts and types but are (1) micrite mud; (2) micrite mud and skeletal detritus; (3) skeletal detritus; and (4) less-than-common-to-rare amounts of pellets. Pellets, very significant in most limestone units of facies previously discussed, here contribute little. This facies can be divided into the following subfacies: (1) mudstone, with small amounts of skeletal debris; and (2) detrital skeletal wackestone. Solitary coral-bearing subfacies are not well developed within this facies. Total coral development formed three very thin coral-bearing beds, with less-than-common-to-rare solitary corals. This facies then has a noted lack of coral development.

Wackestone rock types which form the dominant subfacies contain varying amounts of detrital skeletal debris (fig. 25). In this figure most skeletal detritus is crinoid columnals and fewer brachiopod or pelecypod shells, all recrystallized and set in micrite mud which has been slightly recrystallized. Recrystallization of the micrite mud or small nondescript skeletal detritus began with formation of very small calcspars. Changes in skeletal types became obvious as all the thin sections were studied with occasional shifts of common amounts of crinoid columnals, which make up most of the skeletal detritus, becoming of less importance and not contributing so much detritus to

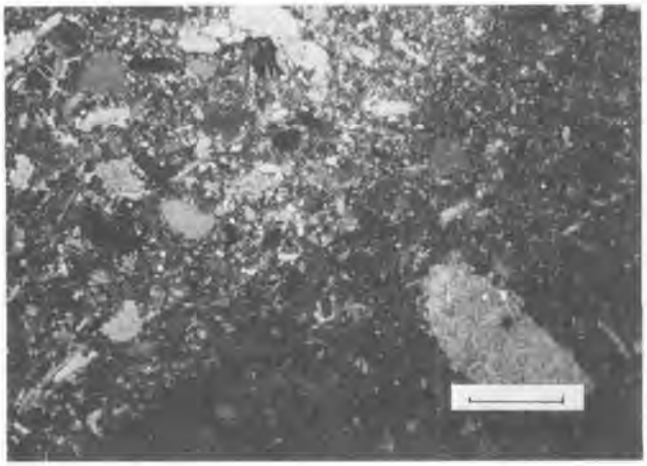


FIGURE 25.—Photomicrograph (crossed nicols) of skeletal wackestone, right side stained with Alizarin Red S. Scale is 1 mm.

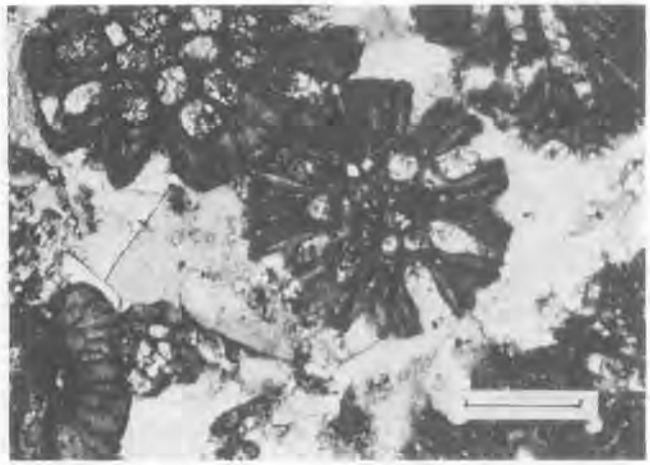


FIGURE 27.—Photomicrograph (crossed nicols) of a bryozoan wackestone. Scale is 0.5 mm.

form some wackestones, with planktonic whole skeletal material becoming more common (fig. 26). In this figure detrital skeletal debris is composed of only a few brachiopod or pelecypod shells with more whole foraminifera skeletons set in slightly recrystallized micrite mud. Most thin sections reveal bryozoans to be highly fragmental with many replaced by silica, but in one unit, 80, conditions of deposition to form a skeletal wackestone must have been such as to deposit whole bryozoans (fig. 27). Here whole cryptostome bryozoans, of the family Rhabdomesidae, are set in recrystallized primary pore cement with smaller calcspars of sparry calcite filling autopores, which are extending directly out of the photomicrograph. All bryozoans (fig. 27) appear to be the genus *Megacanthopora* except the bryozoan to the lower left which is the genus *Nicklesopora.*

Though rare, pellets, at a few levels in this facies, do become the dominant rock-building constituents (fig. 28). In this figure all dark ovate-shaped areas are algal pellets which compose most of the sample, with lesser amounts of small recrystallized skeletal detritus of crinoid columnals, echinoid plates, brachiopod or pelecypod shells, and few whole foraminifera. Lighter areas between pellets and minor skeletal detritus are patches of matrix where micrite mud is partially recrystallized to form small interlocking sparry calcite. Darker areas in the matrix are locations which are more algal enriched and therefore less affected by recrystallization.

Mudstone subfacies occasionally interfingers between skeletal wackestone portions of this facies to form micrite mud-rich portions where skeletal rock-building constituents are deficient (fig. 29). This figure is a typical, very fine-grained mudstone with only one large piece of a detrital crinoid columnal with matrix composed of micrite mud which has experienced only slight recrystallization. The stained right half of the thin section reveals very small unstained areas of clastic-appearing debris, composing less than 5 percent of the sample. Closer examination of clastic debris reveals it to be silica which has replaced skeletal detritus.

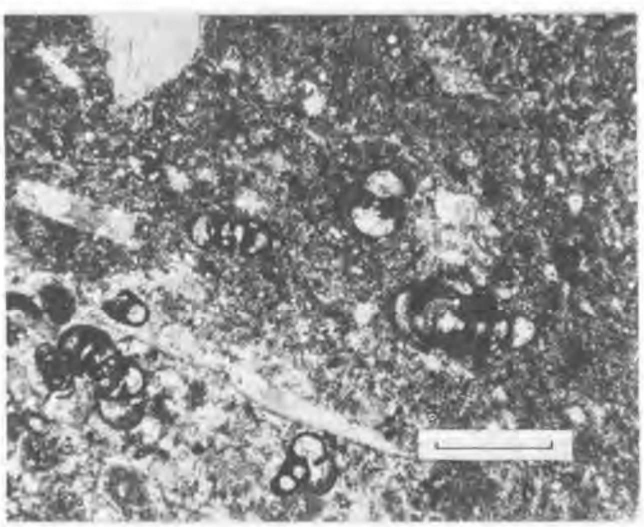


FIGURE 26.—Photomicrograph (crossed nicols) of skeletal wackestone with whole foraminifera present. Scale is 0.5 mm.

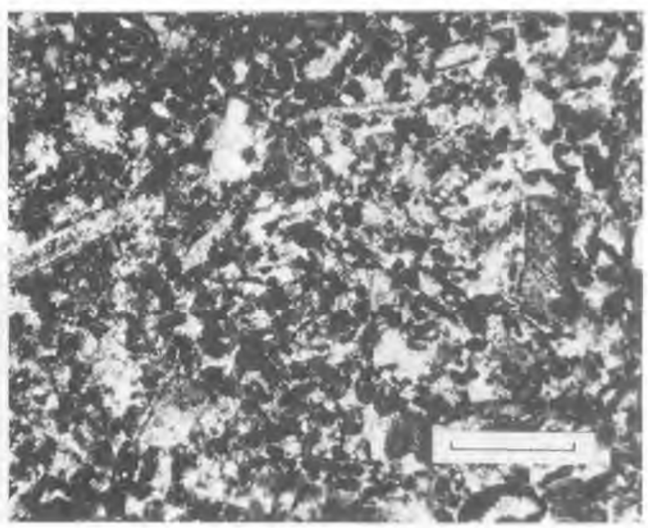


FIGURE 28.—Photomicrograph (crossed nicols) of skeletal, pelletal wackestone. Scale is 0.5 mm.

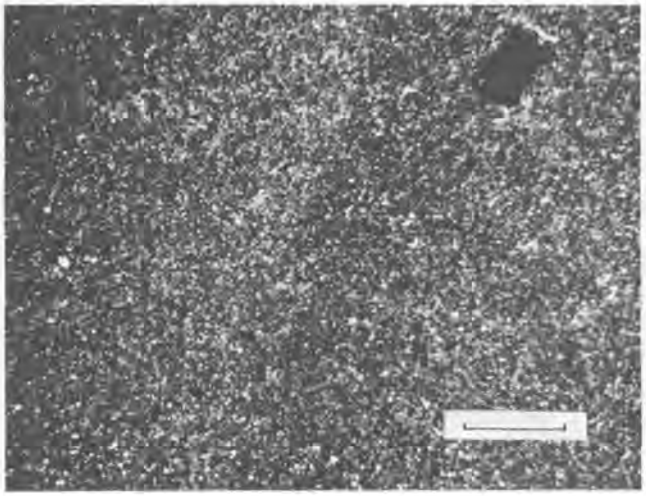


FIGURE 29.—Photomicrograph (crossed nicols) of mudstone with small irregular-shaped silica grains. Right side is stained with Alizarin Red S. Scale is 1 mm.

FIGURE 30.—Photomicrograph (crossed nicols) of mudstone, right side stained with Alizarin Red S. Scale is 0.5 mm.

Mudstone-Packstone Dolomite Sandstone Facies

Uppermost portion of the upper limestone member is characterized by the presence of sandstone, dolomite, and sandy limestone units which interfinger with limestone units to form the last facies in the measured section. This change is encountered in the field by the presence of a basal unit of calcareous sandstone which then gives way upward to varying amounts of intercalated limestone, dolomite, and sandstone sequences. These various intercalations form three subfacies: (1) limestone to sandy limestone, (2) dolomite to sandy dolomite, and (3) sandstone to calcareous dolomitic sandstone. Each subfacies displays abrupt, sharp, complete changes in rock type. A noteworthy characteristic of this facies is its similarity to the first facies forming the base of the Great Blue Formation. A second distinction of it is lack of claystone units which characterize the median shale member.

Limestone rock types form mudstones, wackestones, and packstones, many with clastic admixtures to form sandy limestones. Rock-building constituents of these limestones are (1) micrite mud, (2) skeletal detritus, (3) pellets and skeletal detritus, and (4) clastics. Mudstone becomes the more dominant limestone type when skeletal, pelletal, and clastic constituents become almost negligible (fig. 30). In this figure micrite mud, together with a few very small clusters of sparry calcite, forms a laminated mudstone with a faint lamination shown in the center of the photomicrograph. Higher magnification reveals that some clastics, less than 5 percent, are present; some are rounded, and others are irregularly shaped. They may be released silica particles which had originally replaced skeletal detritus as in the wackestone facies. Packstones predominate as many skeletal rock-building constituents become dominant-to-very-abundant rock constituents (fig. 31). This figure shows a grain-supported skeletal packstone with most skeletal detritus enclosed within algal micrite owing to the debris having been rolled to accumulate an algal micrite outer coating or envelope. Skeletal types are all detrital in nature and are crinoid columnals, echinoid plates, bryozoans, brachiopods, and pelecypods. Lighter areas in the matrix are locations where clear, recrystallized sparry calcite has formed in small interstices between grains, where primary pore cement was recrystallized to form larger interlocking calcspars. Darker areas are slightly recrystallized portions of micrite mud matrix. Grain boundaries have impinged slightly to form welded boundaries between skeletal detritus because of grain-supported conditions. Where clastics interfinger into areas of limestone deposition to form sandy limestone units, it is of interest to note that skeletal detritus and pellets are still the rock-forming constituents together with clastics (fig. 32). In this figure clastics compose 15 percent of the sample with skeletal detritus of crinoid columnals, a few echinoid plates, and brachiopod or pelecypod debris. All dark bodies in the figure are algal pellets and intraclast-bearing algal-coated grains of small skeletal debris.

Dolomite subfacies form five units midway within this facies, from lower calcareous dolomites to upsection thin, sandy dolomites. Units of dolomite or sandy dolomite appear rather abruptly in the section but are not accom-

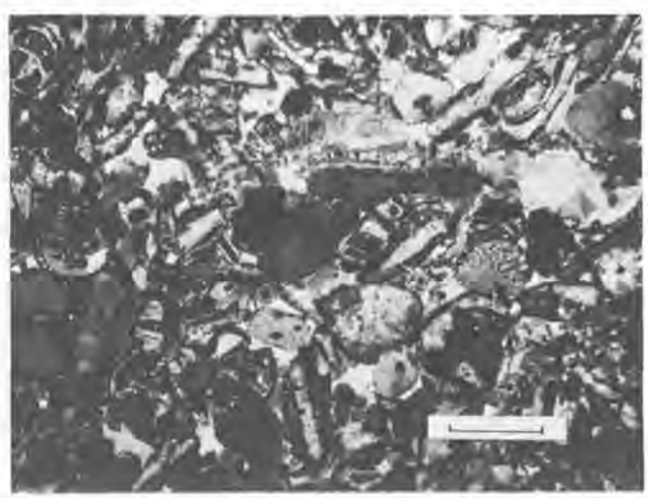


FIGURE 31.—Photomicrograph (crossed nicols) of algal micrite, skeletal packstone. Scale is 1 mm.

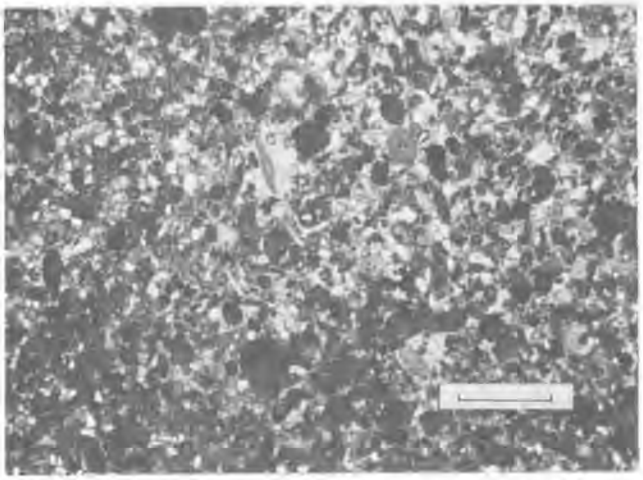


FIGURE 32.—Photomicrograph (crossed nicols) of sandy, pelletal-skeletal wackestone. Scale is 1 mm.

panied by gypsum accumulations, a normal indicator of restricted environment sedimentation (fig. 33). This completely stained thin section reveals a fracture filled with calcite to have taken a dark stain whereas the fine-grained matrix is left mostly unaffected. Here all but a few of the small interlocking grains are subhedra to euhedra of dolorhombs with a few, less than 5 percent, fine clastics scattered within the matrix. Dolorhombs are not well formed because of fine amounts of calcite within the sample which may have slightly dedolomitized any well-developed rhombs which were originally formed. All dolomite and clastic grains are interlocking, with welded grain boundaries and some impingement of grains which embay into sides of other grains.

Clastics at a few levels in this facies accumulated to a point where the clastic compositional fence was crossed to produce sandstone with either calcareous or calcareous dolomitic cementing agents (fig. 34). Most clastics are fine to very fine grained to coarsely silty and are fairly well sorted, with most grains impinging into each other

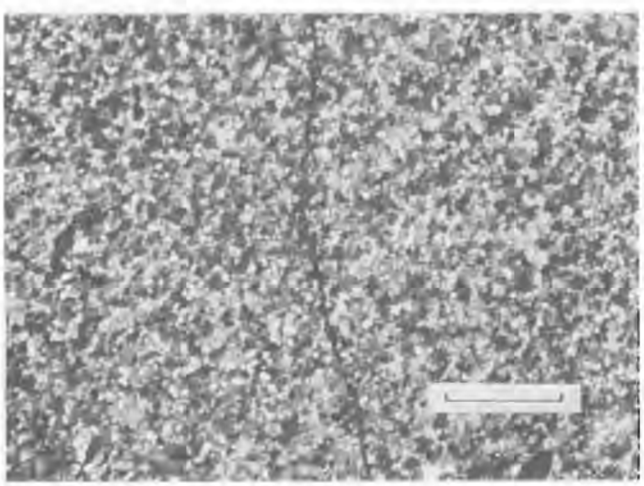


FIGURE 33.—Photomicrograph (crossed nicols) of calcareous dolomite. Sample totally stained by Alizarin Red S, with calcite fracture in center. Scale is 0.5 mm.

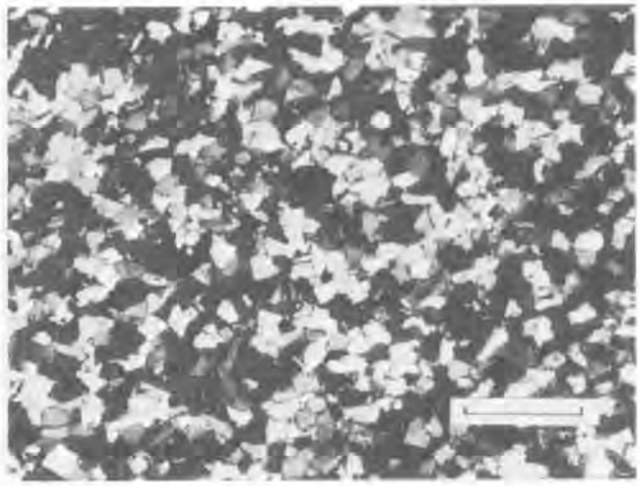


FIGURE 34.—Photomicrograph (crossed nicols) of calcareous sandstone. Scale is 0.5 mm.

to produce embayments into other grains or simply to form welded grain boundaries.

DEPOSITIONAL ENVIRONMENTS

Any attempt to reconstruct ancient depositional environments and develop a working model to explain the presence of seven facies which form Great Blue Formation at Wellsville Mountain necessitates a review of studies which have attempted to reconstruct similar depositional environments, especially those pertaining to formations of Mississippian age. Since 1958 several studies of carbonate depositional environments or generalized environmental reconstructions have been put forth; those consulted include (1) Edie (1958), who discussed Mississippian sedimentation in southeastern Saskatchewan; (2) Shaw (1964), who proposed a general theory of epeiric marine facies relations; (3) Purdy and Imbrie (1964), who described recent carbonate sediments, Grand Bahama Bank; (4) Irwin (1965), who also proposed the general theory of epeiric sea sedimentation; (5) Schenk (1967), who compared Macumber Formation in Mississippian maritime provinces, Canada, to recent strand line development in the Persian Gulf; (6) Laporte (1969), who recognized transgressive carbonate sequences in the lower Devonian Helderberg Group in New York State; (7) Heckle (1972), who described recognition of shallow marine environments; (8) Ryan and Goodell (1972), who studied recent marine geology and estuarine history of Mobile Bay; (9) Rees, Brady, and Rowell (1976), who reconstructed an upper Cambrian carbonate environment; (10) Jehn and Young (1976), who studied Mississippian deposition of the Pitkin Formation in northern Arkansas; (11) Sando (1976), who put forth Mississippian history of the northern Rocky Mountains; and (12) Rose (1976), who also discussed Mississippian carbonate shelf margins of the western United States.

Great Blue Formation is part of an upper Mississippian oscillating epeiric sea, or what Rose (1976) refers to as part of an upper depositional complex which formed a broad carbonate shelf margin (lower Mississippian forming a lower depositional complex). At Wellsville Mountain two marine transgressions with an intervening episode of epeirogeny are recorded in the Mississippian rock sequences. The first transgression deposited Lodgepole Limestone and a thin Deseret Limestone with the second transgression depositing Humbug Formation, Great Blue Formation and Manning Canyon Shale. These are separated by an intervening time of epeirogeny (Sando 1967, 1968, 1975). Shaw (1964) and Irwin (1965) both present similar ways of recognizing environment patterns in rocks deposited in epeiric seas by use of ancient energy levels which affected the broad epeiric sea.

Irwin (1965) recognized three energy zones which he named zones X, Y, and Z. These zones covered very broad areas within epeiric seas and may have extended for from tens to hundreds of kilometers across the broad sea floor. Zone X formed in marine waters deep enough not to have the sediment floor affected by wave action, with very little sediment accumulation. Cooler water temperatures kept waters undersaturated, and so no carbonate precipitates formed. Deeper water also effectively limited sunlight on the sediment floor to restrict photosynthesis and development of algae. Most sediment deposited in zone X was produced in zone Y and transported into this zone. Zone Y was present where waves touched bottom and disturbed the ancient sea floor, with friction essentially dissipating wave action by reducing wave energy. The end of zone Y is where a wave action is effectively stopped. Zone Y could have been tens of kilometers wide and was affected by high amounts of energy to produce ooids and have animal life, including reefs, present on the cratonward side. Zone Z was formed behind zone Y, may have been hundreds of kilometers in width, extending to the strand line. Here less circulation was present because of loss of wave action, but local waves generated by wind, local currents, storms, and probably some tidal changes still affected this zone somewhat. At the strand line evaporites and dolomite developed owing to high rates of evaporation by intense shallowing and ease of heating the marine waters. Seaward side of zone Z interfingered with zone Y to produce reefs with tidal inlets scattering skeletal debris back into zone Z and algal pellets formed slightly farther cratonward behind accumulating skeletal debris. Lagoonal conditions could form between the pelletal zone and the strand line. As these zones migrated back and forth because of slight fluctuations in water depth or subsidence rates, they interfingered various parts of the rock-forming environments to produce an interbedded depositional complex of several smaller environments. Zones Y and Z are recognized at Deweyville section of Wellsville Mountain. As a result of detailed thin-section analysis, field investigations, and review of the literature, the following depositional model for this particular part of the Great Blue depositional complex is herein proposed (fig. 35). Deposition is part of what Sando (1976) refers to as cycle III in Mississippian paleotectonic history.

To provide details of deposition of various units of Great Blue Formation at Wellsville Mountain, in respect to the depositional model (fig. 35), the seven facies present at Deweyville section will be discussed individually to arrive at more specific environments which formed each facies.

Lowermost wackestone dolomite sandstone facies was formed by conditions of deposition somewhat similar to those that formed Humbug Formation and then interfingered with limestone conditions of deposition more characteristic of Great Blue Formation. Clastics, which are very common in Humbug Formation, are only locally characteristic in this lowest facies. No clay minerals or clay-sized clastics were deposited, but silt-sized to very fine-grained sand-sized clastics did accumulate. Much finer clay minerals and clastics representative of an advancing shore were not deposited, possibly because of sedimentary bypassing leaving only slightly coarser clastics. Humbug Formation and parts of lowermost facies of Great Blue Formation received cratonic sands which spread well into the depocenter after broad epeirogeny in latest early Meramecian time had ceased (Sando 1967, 1968, 1975, 1976; Rose 1976). Various sandstone sequences in the basal

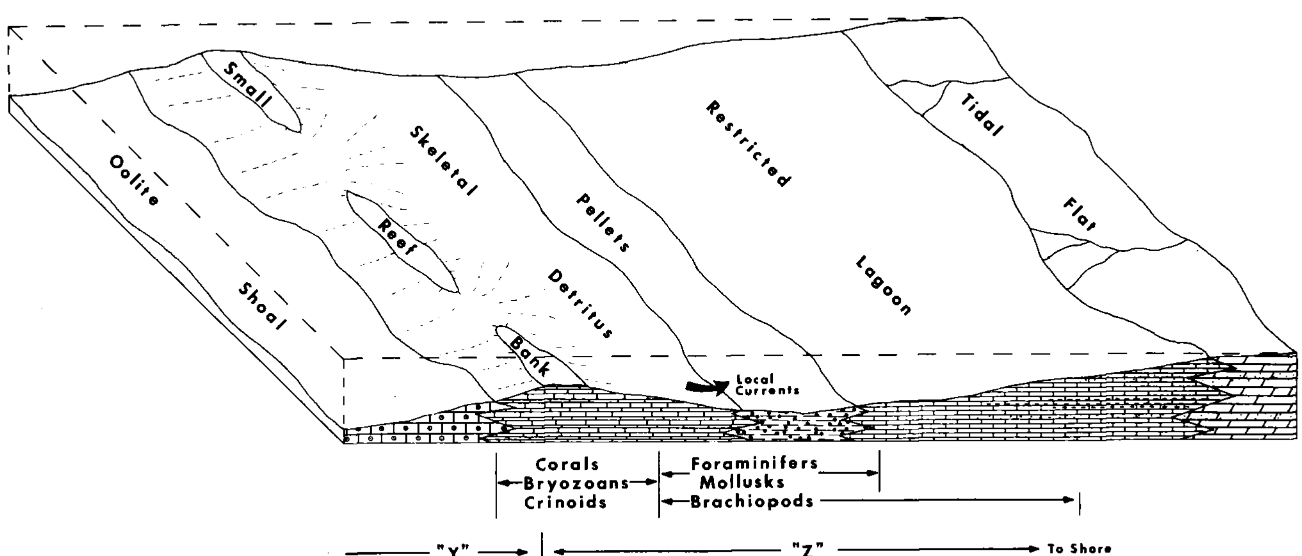


FIGURE 35.—Depositional model for Great Blue Formation at Deweyville section of Wellsville Mountain. Zones Y and Z after Irwin (1965). Skeletal zonation after Jehn and Young (1976).

facies of Great Blue Formation identify localities where a few tongues of clastics were still being shed into the depocenter. Dolomite present in some units in this basal facies is not accompanied by other evaporites, such as gypsum, and must not have formed near the strand line. Adams and Rhodes (1972) propose that dolomite may be formed by seepage refluxion of hypersaline brines into a restricted lagoon-open ocean couplet to alter limestone to dolomite. Chilingar and Bissell (1972), in a discussion of this proposal, believe this to be a viable way of producing dolomite without the necessary presence of other evaporites. Limestone composed of small amounts of clastic and dolomite constituents to purer limestone which becomes rich in detrital skeletal rock-building constituents formed as the environment shifted out of lagoonal conditions into a more open marine limestone-generating subenvironment. Changes in rock types which interfinger vertico-laterally then formed this particular facies within an open-ended, lagoon to shallow marine environment which collected skeletal detritus.

By the time wackestone-packstone facies (second) was reached, clastics and dolomite were absent, with only limestone accumulating. This situation represented a seaward shift of the total environment to form completely in shallow marine conditions. Subfacies of different limestones began to form as deposition slowly shifted or oscillated back and forth between thin oolite-bearing beds to a zone enriched with skeletal detritus transported in from areas of abundant skeletal development. Deposition changed slightly to move out or away from enrichments of skeletal detritus to where a zone rich in pellets and poor in skeletal debris was reached, slightly closer to the craton. This pelletal zone represents the farthest toward the strand this second facies may have reached. Limestone deposition and development of corals, especially a coral biostrome, provide proof of a marine transgression (Link 1975).

Oolite facies (third) represents a basinward shift, the farthest out into the depocenter, forming a fairly continuous sequence of ooid generation. Most ooids have detrital skeletal nuclei and are accompanied by skeletal intraclasts, some of which have algal envelope coatings. Occasionally oolite facies changed slightly to form coral-bearing beds within ooid-bearing beds, with one shift farther back into conditions briefly depositing skeletal debris. This facies does not represent an actual ooid-generating environment but is more representative of a section probably present within a tidal inlet and accounts for both skeletal debris and ooid development.

Mudstone-packstone facies (fourth) developed as the environment shifted back behind zones of ooid and coral generation where accumulations of skeletal detritus and pellets were the dominant rock-building constituents. This facies contains few solitary coral-bearing beds upsection, with deposition mostly of skeletal detritus, pellets, and mudstone of micrite mud. Mudstone may have formed in front of or behind the pelletal zone, possibly in part of an open-ended lagoonal environment. These shifting or oscillating facies then formed between the outer perimeter of a lagoonal environment into a shallow marine environment depositing mudstone, pellets, skeletal detritus, solitary corals, and ooids to produce the lower limestone member.

Median shale member composes all the mudstone-wackestone claystone facies (fifth) and reflects a drastic change in the environment which now received interfingerings of clays and very fine clastics to form claystone units between limestone units. These contrasting rock units represent a shifting environment which began to oscillate between an open-ended lagoon, depositing clays and very fine clastics from an encroaching shoreline, to limestone units in the outer perimeter of the lagoon forming mudstone and occasionally shifting farther out into pelletal and skeletal detritus zones. Micrite nodules within claystone units reveal some fine-grained limestone deposition to have interbedded with clays and clastics in a low-energy, quiet-water environment, still within the lagoon. Ryan and Goodell (1972) have studied Mobile Bay, an estuary connected with the Gulf of Mexico; they found finer clay- to silt-sized material forming farthest down through the bay, with coarser clastics landward toward the delta at the top of the bay. This estuarine bay may serve as a small-scale model for what may have been a much larger lagoonal bay which deposited shale or claystone units throughout the median shale member of Great Blue Formation with interbedded limestone units forming because of shifts in the environment out into the shallow marine pelletal and detrital skeletal environments.

Once the last calcareous shale unit was deposited, limestone sequences again become the only rock type and form the lower portion of upper limestone member. This lower part is composed mostly of wackestone and some mudstone with high amounts of skeletal rock-building constituents and few pellets to form the wackestone facies (sixth). Deposition now shifted from oscillating lagoonal conditions to an open, shallow, marine environment and remained there receiving large amounts of skeletal detritus with lowered influx during some periods. Pellets are formed less often here than in other limestone facies; only rarely then was the pelletal zone part of the environment forming this facies. Skeletal constituents changed from one dominant skeletal group to another. Crinoid, brachiopod, and bryozoan fragments are the most common skeletal group, but occasionally they do shift out of the environment to be replaced by occasional blooms of foraminifera. Solitary corals are rare in the wackestone facies and seem to mark a seaward limitation at which the facies was only rarely formed. To summarize, this facies was formed in a zone receiving skeletal detritus with a seaward limit basically where beginnings of solitary coral occur and a cratonward limit where pellets occur.

Mudstone-packstone dolomite sandstone facies (seventh) was the last in the measured section to be deposited. It is similar to the first, basal facies, being composed of interfingerings of limestone, dolomite, and sandstone with smaller amounts of clastics forming sandy limestone. Limestone sequences characterize deposition which occurred within the skeletal detritus zone and pelletal zone as well as deposition poor in skeletal debris and pellets, forming mudstone which could have been present on either side of the pelletal zone. Limestone forming this facies was then deposited in shallow marine areas behind oolite shoals and very small local reefs or banks which were farther out into the depocenter. Clastics are again present, very similar in occurrence and appearance to clastics within the basal facies, not accompanied by clay minerals and finer clastics. Four distinct occurrences of calcareous sandstone are present and may have been caused by minor uparching occurring within the miogeosyncline. Source areas for clastics may have been from two separate areas. One source area may have been northwestern Utah where uplifts in the Pennsylvanian likely became active in latest Mississippian time (Eardley 1947, 1962). A more logical area for clastic generation may have been to the east and northeast, the Bannock uplift (Sando 1976; Rose 1976), which become active in Late Mississippian. Local currents or storms were then probably responsible for distributing fine-grained to coarsely silty clastics throughout the depocenter to form thin blankets of sand. The five dolomite units in this facies are interpreted as having formed by seepage refluxion of brines out into a lagoonal environment which dolomitized sediments in the lagoon floor. Large amounts of iron in dolorhombs and disseminated throughout dolomite samples, forming iron oxide, may have been produced by reducing conditions within the lagoon. This last facies then accumulated in an environment which oscillated between an outer perimeter of a lagoon and shallow marine waters, producing mudstone, pellets, and skeletal debris.

Top of Great Blue Formation is a slightly covered unconformity with cross-bedded skeletal limestone of Oquirrh Formation above Great Blue Formation. No Manning Canyon Shale was found between the two formations at this locality, indicating a substantial hiatus or disconformity. Latest Mississippian through earliest Pennsylvanian time then marks an interval when uplift affected the area, probably after deposition of Manning Canyon Shale, because of its presence at Dry Lake section of Wellsville Mountain. Whether Bannock uplift or the uplift located in northwestern Utah was responsible for the formation of this unconformity is unknown. Relocation of the Cache allochthon back to its approximate original position, prior to thrust development, places the measured Deweyville section very near the Pennsylvanian uplift in northwestern Utah of Eardley (1947, 1962).

Through the complete Deweyville section most contacts between various units are very abrupt and sharp. In the mudstone-packstone dolomite sandstone facies (seventh), a diastem was located which has silty mudstone with large numbers of chert nodules abruptly overlain by skeletal, sandy, laminated limestone, which cuts through chert nodules in the lower silty mudstone (fig. 36). In this figure is the best example of an oscillating environment which may have abruptly shifted environments within the depositional model. Sharp contacts between unlike units which form total changes in rock type may also be the result of abrupt shifts in the environment. Sandstone units mark very abrupt changes as they overlie limestone units and may be indicative of small-scale hiatus development between the two rock units (Dunbar and Rogers 1957).

FIGURE 36.—Diastem located in mudstone-packstone dolomite sandstone facies. Staff is marked in 1-foot intervals.

Chert was found at only certain elevations in the section and tended to be confined to only certain horizons within individual units. One would surmise that chert deposition, as bedded, nodular, or irregular shaped masses, indicates high concentrations of silica entering the depocenter, but in rather spasmodic, abrupt influxes of siliceous material which then abruptly ceased, only to reappear again later (higher in the section) but in no set pattern. Chert is not very common in the lower limestone member, becomes slightly more common in the median shale member, and is most common in the upper limestone member (fig. 3). Thin sections of chert nodules reveal three dimensionally interlocking submicroscopic silica with only rare sponge spicules present in a few thin sections. Chert deposited within Great Blue Formation could have two distinctively different source areas bringing silica to the depocenter to form the various chert blebs and stringers. Rivers draining the craton could have possibly brought enrichments of silica into the shallow marine environment (Bien, Contois, and Thomas 1959). Another source area was the Antler orogenic belt which was active in Late Mississippian time, producing tremendous quantities of siliceous material (Bissell 1959). Antler orogenic belt may have experienced several minor pulses of volcanism to form alternating periods of addition of high amounts of silica to the miogeosyncline, probably accounting for the episodic appearance of chert in the study area. Moore (1957) states that tremendous amounts of chert which make abrupt appearances and disappearances in a rock column may be directly related to large-scale orogenic pulsations. Two very different but important localities laterally removed from, but flanking, the Great Blue depositional complex on the east and west may have contributed significant quantities of silica to the depocenter.

One characteristic feature of Great Blue Formation is the dark gray color of the limestone units. After a summer rain or upon breaking rocks apart, a fetid odor is released. This fetid odor was also noted during preparation of thin sections. Rock samples which were acidized in dilute HCl acid for approximately 24-48 hours left residues of some silt but mostly a dark-gray-to-black tarlike substance. This dark substance may be iron monosulphide or pyrite produced by slight reducing conditions to form submicroscopic, disseminated particles within the sample (Berner 1963, 1964, 1970). This would indicate that during deposition of Great Blue Formation reducing conditions were active within the depocenter, at least locally. Just how much reduction was taking place is difficult to ascertain. Reducing conditions, as such, would not permit development of skeletal materials, which are the dominant rock-building constituents. Detrital nature of most skeletal constituents, except foraminifera which were probably planktonic, would indicate that skeletal development and growth was taking place elsewhere in the depocenter where better circulation in marine waters was present. Broken-up skeletal detritus was then transported to the study area. Rose (1976) places the crest of the Late

Mississippian upper shelf margin approximately 70 kilometers west of the study area. If this is correct and if amounts of thrust translation of the Cache allochthon of 64 to 160 kilometers is also correct (Crittenden 1961, 1972, pers. comm., Sept., 1976), Deweyville section of Great Blue Formation was located very close to the crest of the shelf margin (fig. 37). Reducing conditions could have then been active behind the crest, with coral and other skeletal development, along with oolites, at or near it. Reduction was then probably taking place in areas within the depocenter which were located behind where ooids and very small coral communities and skeletal constituents were originally developed, in Irwin's (1965) zone Z.

Trace fossils also characterize certain limestone units in Great Blue Formation and appear to have had no one particular grazing path orientation, because tracks were found in horizontal, vertical, and oblique directions. All trace-fossil grazing trails appear to be of the Nereites types (Seilacher 1964; Rhoades 1975). Other fossils present are microfossils of endothyroid foraminifera which were found in thin section. A detailed zonation of them was beyond the scope of this study, but zonations have been accomplished in central Utah by Woodland (1958) and Zeller (1950, 1957) and in central Idaho by Skipp (1961). From their illustrations of endothyroid foraminifera, species of both *Plectogyra* and *Paramillerella* were recognized to be present in various units within the Great Blue Formation.

FIGURE 37.—Reconstruction of upper Mississippian carbonate shelf margin of the western United States, after Rose (1976). Arrow points to study area.

CONCLUSION

Great Blue Formation is herein interpreted to have formed in a shallow marine environment which formed seven very distinct and characteristic facies at Deweyville section of Wellsville Mountain. These seven facies experienced shifts of environments back and forth in a cyclic depositional pattern within the Cordilleran miogeosyncline being formed by the following environments: (1) shoaling ooid-producing waters which formed farthest out into the depocenter; (2) very small localized reefs or banks of mostly solitary corals probably with bryozoans and crinoids, forming behind the shoal; (3) skeletal detritus scattered behind small reefs and banks of coral, crinoid, and bryozoan development, skeletal debris transported possibly through tidal inlets by tidal changes, storms, local currents, etc.; (4) a pelletal-enriched zone farther cratonward where increased amounts of pellets become dominant with decreased or negligible amounts of skeletal debris; and (5) lagoonal conditions behind the pelletal zone nearest the craton. Possible reducing conditions behind the shoal may have been attendant with deposition of some portions of the section to give the rock sequences their characteristic dark color. Chert found sporadically within the section may have been derived from two source areas: (1) streams draining the craton on the east; and (2) volcanic pulsations of the Antler orogenic belt to the west. Shifts within environments of deposition may have been abrupt, leaving sharp contacts between most rock units, with one well-documented diastem present. These continuously shifting or oscillating environments characterize an unstable portion of the eastern part of the Cordilleran miogeosyncline upon which Great Blue Formation accumulated during upper Meramecian and Chesterian time.

(Descriptions of measured sections may be obtained from the author or from Brigham Young University Department of Geology, 259 ESC, Provo, Utah 84602.)

REFERENCES CITED

Adams, J. E., and Rhodes, M. L. 1972, Dolomitization by seepage refluxion: Reprinted from Amer. Assoc. Petrol. Geol. Bull., v. 44, no. 12, p. 1912-1920: Amer. Assoc. Petrol. Geol. Reprint Series 4, p. 135-143.

Armstrong, R. L., 1968, Sevier orogenic belt in Nevada and Utah: Geol. Soc. Amer. Bull., v. 79, p. 429-458.

Berner, R. A., 1963, Distribution and diagenesis of sulfur in some sediments from the Gulf of California: Marine Geol., v. 1, p. 117-140.

———, 1964, Iron sulfides formed from aqueous solution at low temperatures and atmospheric pressure: Jour. Geol., v. 72, p. 293-306.

———, 1970, Sedimentary pyrite formation: Amer. Jour. Sci., v. 268, p. 1-23.

Beus, S. S., 1958, Geology of the northern part of Wellsville Mountain, northern Wasatch Range, Utah: Master's thesis, Utah State University, 84 p.

———, 1963, Geology of the central Bluespring Hills, Utah-Idaho: Ph.D. dissertation, University of Calif., Los Angeles, 233 p.

———, 1968, Paleozoic stratigraphy of Samaria Mountain, Idaho-Utah: Amer. Assoc. Petrol. Geol. Bull., v. 52, no. 5, p. 782-808.

———, 1975, personal communication.

Bien, G. S., Contois, D. E., and Thomas, W. H., 1959, The removal of soluble silica from fresh water entering the sea:

Silica in sediments—a symposium: Soc. Econ. Paleont. Mineral. Spec. Pub. 7, p. 20-35.

Bissell, H. J., 1959, Silica in sediments of the upper Paleozoic of the Cordilleran area: Silica in sediments—a symposium: Soc. Econ. Paleont. Mineral. Spec. Pub. 7, p. 150-185.

———, 1970, Petrology and petrography of lower Triassic marine carbonates of southern Nevada (USA): E. J. Brill, Leiden, Netherlands.

Black, B. A., 1965, Nebo overthrust, southern Wasatch Mountains, Utah: Brigham Young Univ. Geol. St., v. 12, p. 55-90.

Brady, M. J., 1965, Thrusting in the southern Wasatch Mountains, Utah: Brigham Young Univ. Geol. St., v. 12, p. 3-54.

Burchfiel, B. C., and Davis, G. A., 1972, Structural framework and evolution of the southern part of the Cordilleran orogen, western United States: Amer. Jour. Sci., v. 272, p. 97-118.

———, 1975, Nature and controls of Cordilleran orogensis, Western United States: extensions of an earlier synthesis: Amer. Jour. Sci., v. 275-A, p. 363-396.

Butkus, T. A., 1975, Sedimentology and depositional environments of the Great Blue Limestone (Late Mississippian), northcentral Utah: Master's thesis, University of Utah, 142 p.

Carozzi, A. V., and Textoris, D. A., 1967, Paleozoic carbonate microfacies of the eastern stable interior (USA): E. J. Brill, Leiden, Netherlands.

Carr, W. J., and Trimble, D E., 1961, Upper Paleozoic rocks in the Deep Creek Mountains, Idaho: U. S. Geol. Surv. Prof. Paper 424-C (Art. 213), p C181-184.

Chilingar, G V., and Bissell, H. J., 1972, Discussion-dolomitization by seepage refluxion: Reprinted from Amer Assoc. Petrol. Geol. Bull., v. 45, no. 5, p. 679-683: Amer. Assoc. Petrol. Geol. Reprint Series 4, p. 144-146

Crittenden, M. D., Jr, 1959, Mississippian stratigraphy of the central Wasatch and western Uinta Mountains, Utah: Intermountain Assoc. Petrol. Geol. Guidebook 10, p. 63-74

———, 1961, Magnitude of thrust faulting in northern Utah: U. S. Geol. Surv. Prof. Paper 424-D, p. D128-131.

———, 1972, Willard thrust and the Cache allochthon, Utah: Geol. Soc. Amer. Bull., v. 83, p. 2871-2880.

———, 1976, personal communication.

Crittenden, M. D., Jr., Schaeffer, F. E., Trimble, D. E., and Woodward, L. A., 1971, Nomenclature and correlation of some Upper Precambrian and basal Cambrian sequences in western Utah and southeastern Idaho: Geol. Soc. Amer Bull., v. 82, p 581-602.

Dunbar, C. O., and Rogers, J., 1957, Principles of stratigraphy: John Wiley & Sons, New York, 356 p.

Dunham, R. J., 1962, Classification of carbonate rocks according to depositional texture: Classification of carbonate rocks —a symposium· Amer. Assoc Petrol. Geol. Memoir 1, p. 108-121.

Eardley, A. J, 1947, Paleozoic Cordilleran geosyncline and related orogeny: Jour. Geol., v. 55, no. 4, p. 309-342.

———, 1962, Structural geology of North America: Harper and Row, New York, 743 p.

Edie, R. W., 1958, Mississippian sedimentation and oil fields in southeastern Saskatchewan: Amer. Assoc. Petrol Geol Bull., v. 42, no 1, p. 94-126.

Eriksson, Y., 1960, Geology of the upper Ogden Canyon, Weber County, Utah: Master's thesis, University of Utah, 55 p.

Foutz, D R., 1966, Stratigraphy of the Mississippian system in northeastern Utah and adjacent states: Ph.D dissertation, Washington State University, 218 p.

Friedman, G. M, 1959, Identification of carbonate minerals by staining methods: Jour Sed. Petrol., v. 29, no. 1, p. 87-97.

Gelnett, R. H., 1958, Geology of southern part of Wellsville Moutain, Wasatch Range, Utah: Master's thesis, Utah State University.

Gilluly, J., 1932, Geology and ore deposits of the Stockton and Fairfield quadrangles, Utah: U. S. Geol. Surv. Prof. Paper 173, 171 p.

Goddard, E N., Trask, P. D, DeFord, R. K., Rove, O N., Singewald, J. T., Jr., and Overbeck, R M., 1970, Rock-color chart: Huyskes-Enschede, 11 p.

Heckle, P. H, 1972, Recognition of ancient shallow marine environments: Recognition of ancient sedimentary environments—a symposium: Soc. Econ. Paleont. Mineral. Spec. Pub. 16, p. 226-276.

Horowitz, A. S., and Potter, P. E., 1971, Introductory petrography of fossils: Springer-Verlag, New York, 302 p.

Irwin, M. L., 1965, General theory of epeiric clear water sedimentation: Amer. Assoc Petrol. Geol. Bull., v. 49, no. 4, p. 445-459.

Jehn, P. J., and Young, L. M., 1976, Depositional environments of the Pitkin Formation, northern Arkansas: Jour. Sed. Petrol., v. 46, no. 2, p. 377-386.

Laporte, L. F., 1969, Recognition of a transgressive carbonate sequence within an epeiric sea: Helderberg Group (lower Devonian) of New York State: Depositional environments in carbonate rocks—a symposium: Soc. Econ. Paleont. Mineral. Spec. Pub. 14, p. 98-118.

Leighton, M. W, and Pendexter, C., 1962, Carbonate rock types. Classification of carbonate rocks—a symposium: Amer. Assoc. Petrol. Geol. Memoir 1, p. 33-61.

Link, T. A., 1975, Theory of transgressive and regressive reef (bioherm) development and origin of oil: Reprinted from Amer. Assoc. Petrol. Geol Bull., v. 34, no. 2, p. 263-294; Carbonate Rocks III: Organic Reefs: Amer. Assoc. Petrol. Geol. Reprint Series 15, p. 49-80.

Majewske, O. P., 1969, Recognition of invertebrate fossil fragments in rocks and thin sections: E. J. Brill, Leiden, Netherlands.

Mansfield, G. R., 1927, Geography, geology, and mineral resources of parts of southeastern Idaho: U. S. Geol. Surv. Prof. Paper 152, 409 p.

Moore, R. C., 1957, Mississippian carbonate deposits of the Ozark region: Regional aspects of carbonate deposition—a symposium: Soc. Econ. Paleont Mineral. Spec. Pub. 5, p. 101-124.

Morris, H T., and Lovering, T. S, 1961, Stratigraphy of the East Tintic Mountains, Utah. U. S Geol. Surv. Prof. Paper 361, 145 p.

Mullens, T. E., and Izett, G. A, 1964, Geology of the Paradise quadrangle, Cache County, Utah: U. S. Geol. Surv. Bull. 1181-S, p. S1-32.

Nygreen, P. W., 1958, The Oquirrh Formation stratigraphy of the lower portion in the type area and near Logan, Utah: Utah Geol. Mineral. Surv. Bull. 61, p. 5-67.

Olson, R. H, 1956, Geology of the Promontory Range: Guidebook to the geology of Utah, no. 11, Geology of parts of northwestern Utah, p. 41-75.

———, 1960, Geology of the Promontory Range, Box Elder County, Utah· Ph.D. dissertation, University of Utah

Parks, J. M., 1951, Corals from the Brazer Formation (Mississippian) of northern Utah· Jour. Paleont, v. 25, no. 2, p. 171-186

Poole, F. G., 1974, Flysch deposits of Antler foreland basin, western United States· Tectonics and sedimentation—a symposium: Soc Econ. Paleont Mineral Spec. Pub. 22, p. 58-82.

Purdy, E. G, and Imbrie, J., 1964, Carbonate sediments, Great Bahama Bank· Guidebook for field trip no. 2 Geol. Soc. Amer. Convention, Nov. 1964

Rees, M N., Brady, M. J., and Rowell, A. J., 1976. Depositional environment of the upper Cambrian Johns Wash Limestone (House Range, Utah) · Jour Sed Petrol., v. 46, no. 1, p. 38-47.

Rhoades, D C., 1975, The paleoecological and environmental significance of trace fossils: in Frey, R. W. (ed.), The study of trace fossils: Springer-Verlag, New York, p. 147-160.

Richardson, G. B, 1913, The Paleozoic section in northern Utah: Amer. Jour. Sci., v 186, p. 406-416.

———, 1941, Geology and mineral resources of the Randolph quadrangle, Utah-Wyoming: U. S Geol. Surv. Bull. 923, 54 p.

Rigby, J. K., 1959, Upper Devonian unconformity in central Utah· Geol. Soc. Amer. Bull., v. 70, p. 207-218.

———, 1976, personal communication.

Rigo, R J., 1968, Middle and upper Cambrian stratigraphy in the autochthon and allochthon of northern Utah: Brigham Young Univ. Geol. St, v. 15, pt. 1, p. 31-66

Rose, P. R., 1976, Mississippian carbonate shelf margins, western United States: Jour. Res. U. S. Geol. Surv., v. 4, no. 4, July-Aug., p 449-466.

Ryan, J. J., and Goodell, H. G, 1972, Marine geology and estuarine history of Mobile Bay, Alabama, part 1, contemporary sediments: Geol. Soc. Amer. Memoir 133, p. 517-554.

Sadlick, W., 1955, The Mississippian-Pennsylvanian boundary in northeastern Utah: Master's thesis, University of Utah, 77 p.

Sando, W. J., 1967, Mississippian depositional provinces in the northern Cordilleran region: U. S. Geol. Surv. Prof. Paper 575-D, p. D29-38.

———, 1968, Regional Devonian and Mississippian stratigraphy in northern Cordilleran region: U. S. Geol. Surv. Prof. Paper 600-A, Summary of paleontologic investigations geological survey research 1968, p. 6.

———, 1975, Diastem factor in Mississippian rocks of the northern Rocky Mountains: Geology, p. 657-660.

———, 1976, Mississippian history of the northern Rocky Mountains region: Jour. Res. U.S. Geol. Surv., v. 4, no. 3, p. 317-338.

Sando, W. J., Dutro, J. T., and Gere, W. C., 1959, Brazer dolomite (Mississippian), Randolph quadrangle, northeast Utah: Amer. Assoc. Petrol. Geol. Bull., v. 43, no. 12, p. 2741-2769.

Schenk, P. E., 1967, The Macumber Formation of the maritime provinces, Canada—a Mississippian analogue to recent strand-line carbonates of the Persian Gulf: Jour. Sed. Petrol., v. 37, no. 2, p. 365-376.

Seilacher, A., 1964, Biogenic sedimentary structures: in Imbrie, J., and Newell, N., Approaches to paleoecology: John Wiley and Sons, New York, p. 296-316.

Shaw, A. B., 1964, Time in stratigraphy: McGraw-Hill, New York, 365 p.

Skipp, B. A. L., 1961, Stratigraphic distribution of endothyrid foraminifera in Carboniferous rocks of the MacKay quadrangle, Idaho: U. S. Geol. Surv. Prof. Paper 424-C, p. C239-244.

Spurr, J. E., 1895, Economic geology of the Mercur mining district, Utah: U. S. Geol. Surv. 16th Ann. Rep., p. 343-455.

Swede, A., 1975, personal communication.

Tower, G. W., Jr., and Smith, G. O., 1898, Geology and mining industry of the Tintic district, Utah: U. S. Geol. Surv. 19th Ann. Rep., p. 601-785.

Williams, J. S., 1943, Carboniferous formations of the Uinta and northern Wasatch Mountains, Utah: Geol. Soc. Amer. Bull., v. 54, p. 591-624.

———, 1958, Geologic atlas of Utah, Cache County: Utah Geol. Mineral. Surv. Bull. 64, 98 p.

———, 1963, Carboniferous and Permian rocks of Utah: Paper 10, Oil and gas possibilities of Utah, re-evaluated: Utah Geol. Mineral. Surv. Bull. 54, p. 89-107.

Williams, J. S., and Yolton, J. S., 1945, Brazer (Mississippian) and lower Wells (Pennsylvanian) section at Dry Lake, Logan Quadrangle, Utah: Amer. Assoc. Petrol. Geol. Bull., v. 29, no. 8, p. 1143-1155.

Woodland, R. B., 1958, Stratigraphic significance of Mississippian endothyroid foraminifera in central Utah: Jour. Paleont., v. 32, no. 5, p. 791-814.

Zeller, E. J., 1950, Stratigraphic significance of Mississippian endothyroid foraminifera: Univ. Kansas Paleo. Contrib. Protozoa, Article 4, p. 1-23.

———, 1957, Mississippian endothyroid foraminifera from the Cordilleran geosyncline: Jour. Paleont., v. 31, no. 4, p. 679-704.