Evidence for two pulses of glaciation during the late Proterozoic in northern Utah and southeastern Idaho

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ABSTRACT

A record of glaciation during late Proterozoic time is preserved in a number of localities extending from the Sheeprock Mountains, Utah, to Pocatello, Idaho, and from the Park City area 40 km east of Salt Lake City to the Deep Creek Range along the Utah-Nevada line. Over much of this area, the glacial deposits and associated rocks thicken westward and form the basal part of a miogeoclinal wedge that accumulated near the late Proterozoic and early Paleozoic continental margin. In the east, such deposits are thin and rest on Archean basement or rocks of Proterozoic Y age; in the west, they are part of thicker sequences in which deposition apparently continued without significant interruption from late Proterozoic into Cambrian time. In many places, the original continuity between the western and eastern parts of the depositional wedge has been obscured by thrusting of Cretaceous and early Tertiary age that carried the thick basinal sequences eastward over those deposited on the continental platform. Recent mapping of Fremont Island in Great Salt Lake, the Wasatch Range between Ogden and Brigham City, and the Sheeprock Mountains shows that glacial episodes represented either by diamictite or by dropstones enclosed in fine-grained laminated beds are separated by as much as 1,000 m of non-glacial deposits, including black slate, alternating graywacke and siltstone, quartzite, and conglomerate. Using reasonable sedimentation rates for such deposits and by comparison with modern analogues, we infer that two episodes of glaciation, each probably consisting of multiple advances and retreats, were separated by a non-glacial interval of a few hundred thousand to a few million years' duration.

Correlation of the allochthonous, miogeoclinal glacial deposits with the single glacial unit present in autochthonous and parautochthonous platform sites is uncertain, but our interpretation of sedimentary facies and paleogeography suggests that only the younger of the two episodes recorded in the allochthon is represented by the diamictites of the autochthon.

*Deceased, November 1982.

INTRODUCTION

Since Hintze (1913) and Blackwelder (1910, 1925, 1932) first described them, the distinctive boulder-bearing black diamictites of northern Utah and southeastern Idaho have been ascribed alternately to glacial (Calkins and Butler, 1943; Crittenden and others, 1952) and non-glacial processes (Condie, 1967; Schermerhorn, 1974). The most recent studies by Varney (1976), Blick (1979), and Ojakangas and Matsch (1980) have provided definitive evidence of glacial activity in the form of striated and faceted clasts, dropstones, isolated sand and gravel clots, and a striated glacial pavement. Although interpretations differ as to the details of paleogeographic setting and depositional mechanisms, most authors now agree that much of the diamictite was deposited in a glaciomarine environment. Similar rocks constitute the earliest deposits of the late Proterozoic and early Paleozoic miogeocline in many parts of the Cordillera (Crittenden and others, 1972; Stewart, 1972; Christie-Blick and others, 1980).

The diamictites are present in many areas of northern Utah and southeastern Idaho (Fig. 1), and it was evident some time ago (Crittenden and others, 1971) that the distribution of these and overlying Paleozoic rocks has been modified by tectonic transport that took place during the development of the Cordilleran fold belt (King, 1969). As a result, the rocks exposed in eastern autochthonous structural settings (center right edge in Fig. 1) are parts of thin and locally deposited sequences in which multiple unconformities attest to interruptions of sedimentation, whereas most occurrences in the allochthon (A in Fig. 1) are parts of thicker sequences in which deposition apparently continued without major tectonic disturbance into Cambrian time. In the autochchthon, glacial sediments[1] accumulated above thick trough deposits of Proterozoic Y age or were deposited directly on crystalline basement of Archean and early Proterozoic age. In the allochthon, similar and apparently correlative glacial deposits are intercalated with thick non-glacial sediments and locally with submarine volcanic rocks. In many areas, the base of these allochthonous sections either is not exposed or is cut out by thrusts, but locally these rocks unconformably overlie units of Proterozoic X age not present within the autochthon.

In the autochthon, and in some areas of the Charleston-Nebo allochthon (in American Fork Canyon, for example) only one unit of glacially related rocks has been recognized, although there is evidence for repeated fluctuations of the ice margin during over-all retreat (Blick, 1979; Christie-Blick, 1980a). In several areas in the allochthon, however, two glacial episodes can be recognized, separated by a sufficient thickness of non-glacial deposits to suggest that the two episodes were separated by a time interval perhaps as long as the Pleistocene. This paper presents the evidence for

[1] In this paper, "glacial sediments" are ones that accumulate in a glacial environment as defined by Boulton and Deynoux (1981). The term "non-glacial" is applied to sediments in which no glacial influence has been detected, even though some may have been derived by glacial activity or may have accumulated in an environment appreciably influenced by glacial meltwater.

Geological Society of America Bulletin, v. 94, p. 437–450, 8 figs., April 1983.

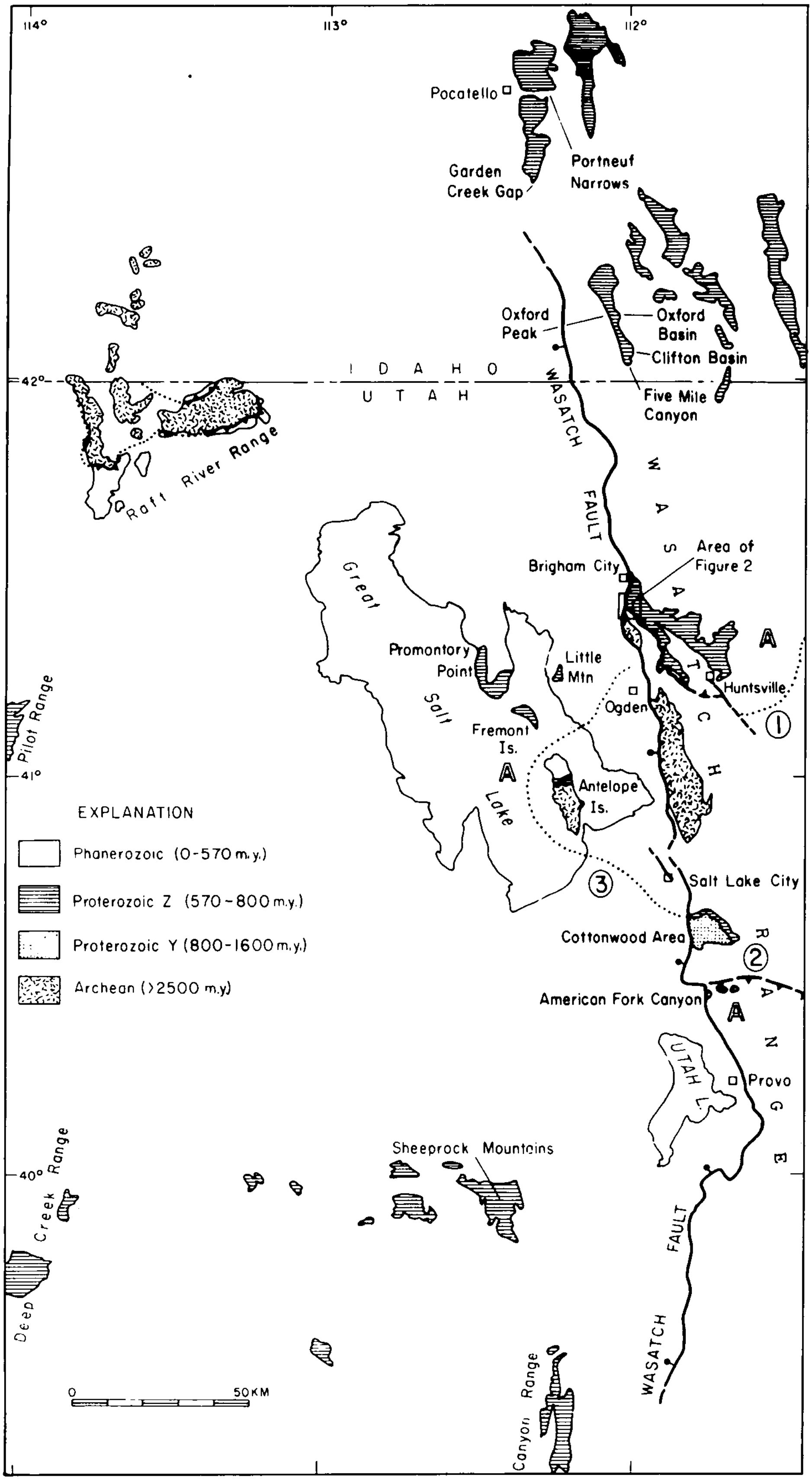


Figure 1. Map of northern Utah and southeastern Idaho showing extent of Precambrian rocks and localities referred to in text. (1) Willard-Paris thrust; (2) Charleston-Nebo thrust; (3) concealed trace of inferred connection between 1 and 2; (A) area of allochthon referred to in text.

this conclusion and assesses its significance for regional correlation of the upper Proterozoic rocks.

GLACIAL RECORD IN THE AUTOCHTHON

Antelope Island

Antelope Island (Fig. 1) is underlain mainly by granite gneiss inferred to be of Archean or early Proterozoic age (Bryant, 1980; Bryant and Graff, 1980; Hedge and others, 1983) and assigned to the Farmington Canyon Complex (Eardley and Hatch, 1940a). These crystalline rocks are overlain unconformably by little-metamorphosed rocks of Proterozoic Z age, consisting of 45 m of diamictite, 40 m of pinkish-tan-weathering laminated dolomite, and 35 m of argillite (only partly exposed). These units are in turn overlain disconformably by an indeterminate thickness, at least several tens of metres, of light tan to pale gray cobble and pebble conglomerate and quartzite, representing the basal part of the Tintic Quartzite (Cambrian). That coarse-grained quartzite was, we believe, correctly identified as Cambrian by Eardley and Hatch (1940b) but later was assigned to the Precambrian Mutual Formation by Larsen (1957), and this assignment was adopted for the geologic map of Utah by Stokes (1963). More recent comparisons with units exposed in the Wasatch Range in both .he allochthon and autochthon confirm Eardley's identification as the Tintic Quartzite and emphasize a close relationship to the exposures in the Farmington-Bountiful section of the Wasatch Range (between Salt Lake City and Ogden) (Fig. 1), where the Tintic Quartzite rests directly on the Farmington Canyon Complex without any intervening rocks of Proterozoic Z age (Crittenden, 1972). These relations indicate that the line of zero thickness for Proterozoic Z rocks lies between Antelope Island

and the Farmington-Bountiful section of the Wasatch.

The tectonic position of Antelope Island is uncertain. Rocks exposed there are structurally beneath the Willard thrust (Fig. 1; Crittenden, 1972), but they may be structurally above several thrusts that crop out in southwestern Wyoming (see Royse and others, 1975; Blick, 1979). We therefore regard Antelope Island as autochthonous or parautochthonous.

Mineral Fork Area

Glacial rocks older than the Cambrian Tintic Quartzite were recognized in the Cottonwood area (Fig. 1) by Hintze (1913) and by Calkins and Butler (1943) during mapping of the Alta and American Fork mining districts. Prior to 1952, the glacial units were called simply "tillite" and, because of the absence of clasts of the distinctive diamictite in the base of the overlying quartzite, were questionably assigned to the Cambrian (Calkins and Butler, 1943). Further mapping (Crittenden and others, 1952), however, showed that the diamictite, there named the Mineral Fork Tillite, was deposited above a major unconformity in two broad, shallow glacial valleys cut in the top of the older Big Cottonwood Formation and was overlapped unconformably by both the Mutual Formation (Proterozoic Z) and the basal conglomerate of the Tintic Quartzite (Early? and Middle Cambrian; see col. 9 of Fig. 4 below). These relations made it clear that the diamictite is separated from the overlying basal Cambrian strata by not merely one unconformity but two. Calkins's observation that clasts of diamictite are absent from the base of the Tintic Quartzite is now seen to be of no age significance.

The sedimentary environment of these glacial deposits (referred to as the Mineral Fork Formation by Christie-Blick) has been discussed most recently by Varney (1976), Ojakangas and Matsch (1980, 1982), Blick (1979), Christie-Blick (1980a, 1982a) and Knoll and others (1981). Evidence of glaciation includes a few dropstones, clots of sand or gravel enclosed in finer-grained strata perhaps analogous to those described by Ovenshine (1970), and striae on the underlying surface (Blick, 1979; Ojakangas and Matsch, 1980). There is considerable disagreement in detail about the interpretation of the Mineral Fork Tillite, but most recent workers agree that it accumulated at or near the margin of an ice sheet that reached the sea. Christie-Blick (1980a) cited evidence of repeated glacial fluctuations during over-all retreat but did not find evidence of prolonged periods of entirely non-glacial deposition. An obvious problem is that in the postulated glaciomarine environment the time interval represented by any particular hiatus is difficult to evaluate. The most that can be said is that there is no evidence that early glacial deposits within the Mineral Fork had consolidated sufficiently to yield clasts of diamictite to later glacial advances.

GLACIAL RECORD IN THE ALLOCHTHON

Mouth of Perry Canyon

In the northern Wasatch Range (Fig. 2), the upper plate of the Willard thrust is characterized by a thick sequence of little-metamorphosed rocks of late Proterozoic age (Sorensen and Crittenden, 1976), apparently recognized by Blackwelder as early as 1910 (Blackwelder, 1910, 1932). Mapping prior to 1970 (Crittenden and others, 1971) showed that these sections, informally referred to as the Huntsville sequence, constitute a consistent set of formational units that are recognizable throughout western Utah and southeastern Idaho and that a diamictite-bearing unit occurs in the lower part of this sequence in most places where the base is exposed. More recent mapping (Sorensen and Crittenden, 1976) has shown that diamictite-bearing rocks, informally designated the "formation of Perry Canyon," are exposed almost continuously above the Willard thrust from the Wasatch front near Perry Canyon at least to Pineview Reservoir west of Huntsville (Fig. 1), a distance of 25 km. This mapping also showed that a higher strand of the thrust repeats both the diamictite-bearing units and the underlying Facer Formation. Inasmuch as the thrust descends stratigraphically westward, the thickest sections of both formations are exposed near the mountain front, where they are cut off by the Wasatch fault. Assuming that this descent continues, the rocks of the upper plate are presumed to have been deposited west of those of the lower plate, but the distance that originally separated them is uncertain; an estimate of 10 to 20 km appears reasonable on the basis of change in thickness of the Facer Formation and other evidence of the total transport on the Willard thrust (Crittenden, 1961; Blick, 1979).

The stratigraphically lowest and thickest section of the formation of Perry Canyon is in the upper plate of the Willard thrust, beginning just south of the mouth of Perry Canyon (Fig. 2), where the formation consists of three units. The lower and upper units are composed of diamictite and the intervening unit is interbedded graywacke and siltstone (see column 5 of Fig. 4 below). The lower diamictite, here about 365 m thick, rests directly on the Facer Formation. Its thickness is variable along strike, however; only 2 km to the east, it is a maximum of 60 m thick and is locally absent. Where diamicite is missing, the base of the Perry Canyon is marked by 1 to 5 m of coarse arkosic grit that grades upward into alternating graywacke and siltstone. Much of the lower diamictite appears massive, and some may be lodgement tillite (see Boulton and Deynoux, 1981), but intercalated lenses of siltstone 2 to 3 m thick indicate intermittent intervals of non-glacial deposition. Clasts are predominantly 5 to 10 cm in diameter, but a few are as large as 1 m and the largest observed is nearly 3 m long. A count of 100 clasts showed gneissic granite (42%) and pale quartzite (34%) to be the most abundant. Clasts of pale metarhyolite (10%) are an unusual component seen only in this area; clasts of schist, basalt(?), dark quartzite, and carbonate rocks make up the remaining 14%. Thin pods of badly altered metavolcanic rocks are present locally at the base of this lowest diamictite. At the top, it is overlain abruptly by graywacke and siltstone, although all of these rocks intertongue locally.

The medial part of the formation of Perry Canyon here consists of interbedded greenish-tan graywacke and dark gray to greenish-gray siltstone. Locally, thin lenses of laminated gray limestone are present at or just above the lower contact. On the ridge south of the mouth of Perry Canyon, lenticular beds of tan-weathering quartzitic graywacke or quartzitic sandstone attain a thickness of 1 to 2 m. To the southeast, the medial unit forms much of the ridge crest between Willard and Perry Canyons. In this area, the dominant lithology is dark gray–to olive-drab–weathering siltstone in which even parallel bedding 3 to 5 cm thick combines with weak foliation, intersecting at an angle of 30° to 40°, to produce a distinctively banded phyllite. Although poorly

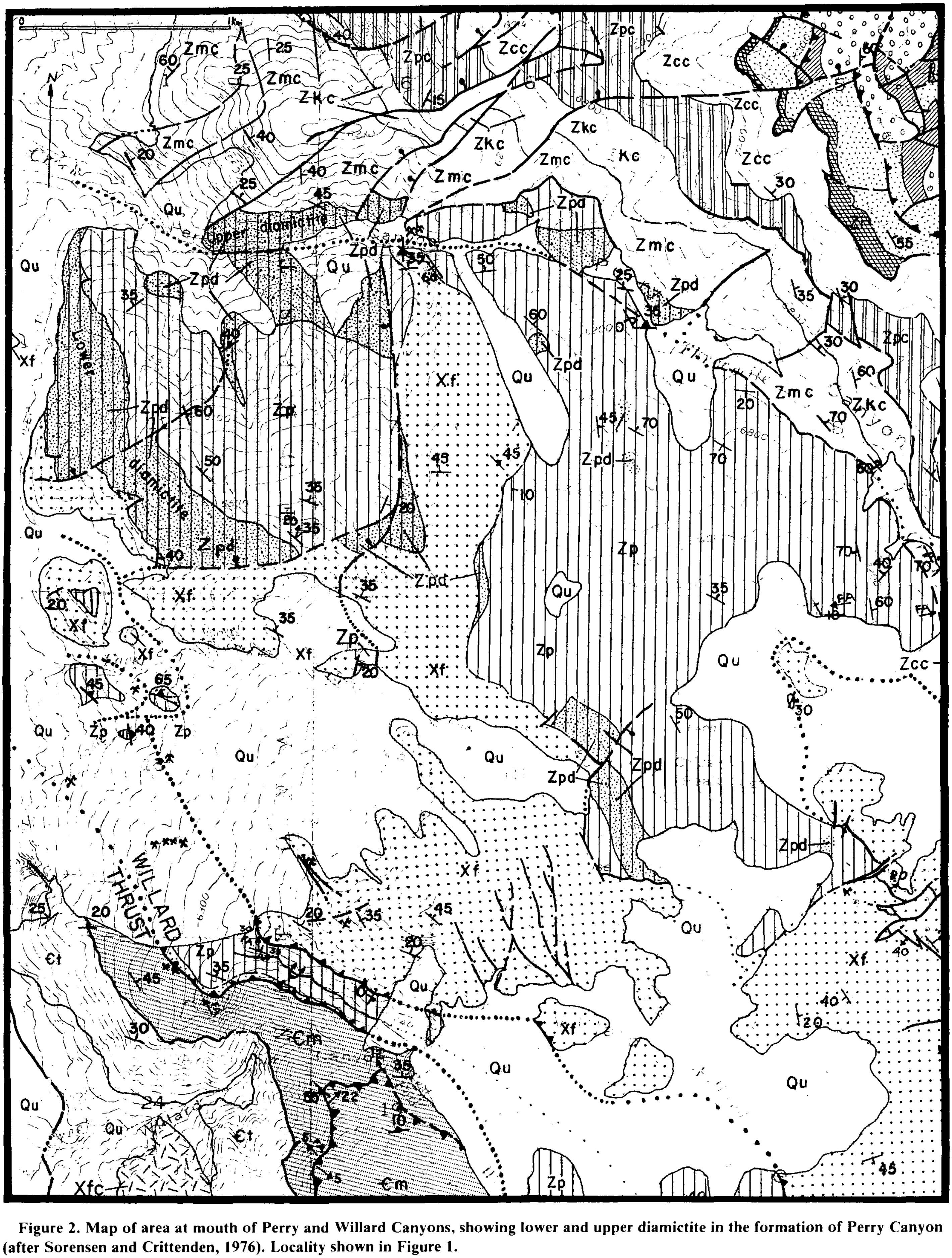
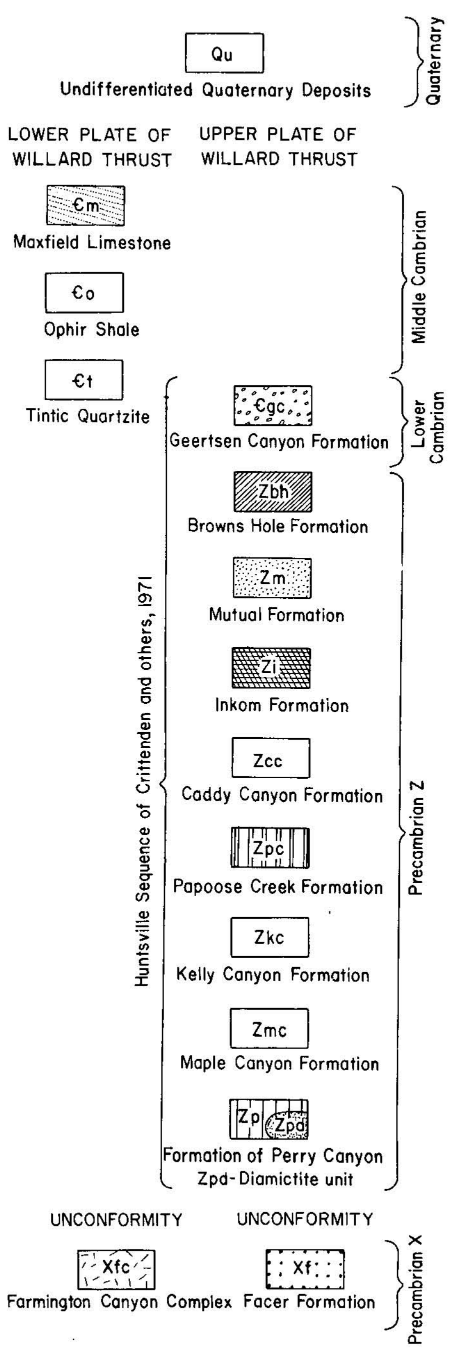


Figure 2. Map of area at mouth of Perry and Willard Canyons, showing lower and upper diamictite in the formation of Perry Canyon (after Sorensen and Crittenden, 1976). Locality shown in Figure 1.

FIGURE 2

EXPLANATION



exposed on the wooded south slope of Perry Canyon, the medial unit appears from a cross section to be about 500 m thick.

The upper part of the formation of Perry Canyon consists of massive diamictite that forms prominent dark cliffs on the north slope of Perry Canyon about 1 km east of the mountain front. Although the base is cut by a fault in Perry Canyon and may actually be in a separate thrust sheet, this body of diamictite appears to overlie the medial unit of graywacke and siltstone and is overlain by laminated greenish-gray, fine-grained siltstone assigned to the Maple Canyon Formation (Crittenden and others, 1971). About 200 m of diamictite is exposed in the cliffs, and an additional thickness of as much as 100 m is exposed on the south slope of the canyon. The total thickness is approximate because of the uncertain effects of the fault that has been mapped following the bottom of the stream canyon to the east. The rocks above and below the two diamictites, however, are quite different, and it is unlikely that they represent a single horizon repeated by faulting. A clast count in the upper diamictite recorded 70% granite and gneissic rocks, 27% quartzite, and about 3% other sedimentary and igneous rocks.

Willard Canyon to Pineview Reservoir

On the north rim of Willard Canyon, a short distance southwest of the exposures just described, the formation of Perry Canyon forms a second band of outcrops that extends almost continuously southeast to Pineview Reservoir about 5 km west of Huntsville (Fig. 1). These exposures are also allochthonous but are in a structurally lower plate of the Willard thrust.

In contrast to the section at the mouth of Perry Canyon, this section contains diamictite only in lenticular pods ranging from 20 or 30 m to perhaps 400 m thick (see columns 6 and 7 of Fig. 4 below). In general, this section shows more evidence of downslope movement of glacially derived sediment (sliding and sediment gravity flow). In addition, submarine volcanic rocks, including pillow basalt (Fig. 3), and shallow intrusive rocks are more abundant, and a 1- to 3-m-thick bed of carbonate that consists of either gray laminated limestone or, more rarely, pinkish-tan laminated dolomite forms a "cap" over several of the diamictite lenses similar to those described by Williams (1979) in Australia. Such caps locally extend beyond the limits of the diamictite lens. Although nowhere do two lenses of diamictite appear one above another, as they do at the mouth of Perry Canyon, some lenses rest directly on the underlying Facer Formation or on the Willard thrust, whereas others are entirely surrounded by non-glacial(?) deposits. Where diamictite is absent, the basal bed usually is a coarse arkosic grit, rarely more than 1 m thick, consisting of angular to subrounded grains in a silty matrix. In many places, it is faulted out or obscured by float. Between exposures of diamictite, the formation consists of dark gray to black, thin-bedded pyritic siltstone intercalated with beds of graywacke, dark gray sandstone, or locally tan-weathering quartzite. Pyrite is disseminated in many of these rocks as cubes as large as 1 cm across. To the southeast beyond North Odgen Pass (Fig. 4), lenses of diamictite are

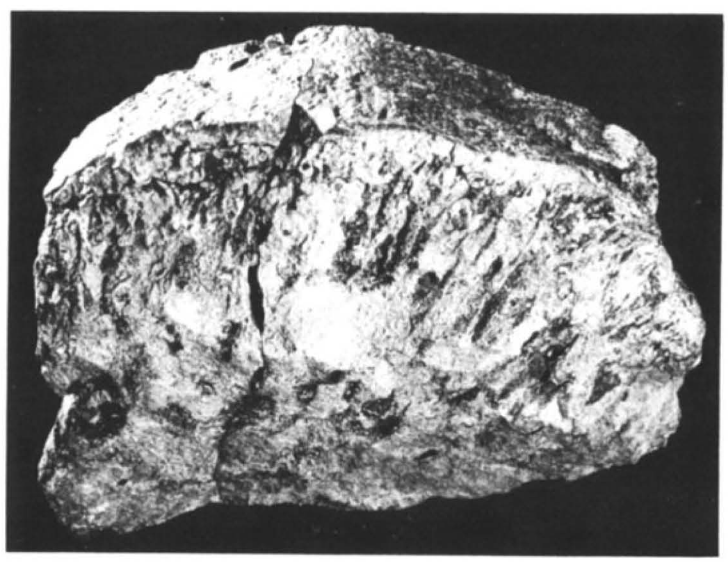


Figure 3. Basaltic pillow from volcanic rocks intercalated with diamictite member of the formation of Perry Canyon, North Ogden Pass, Utah. Specimen is 14 cm long.

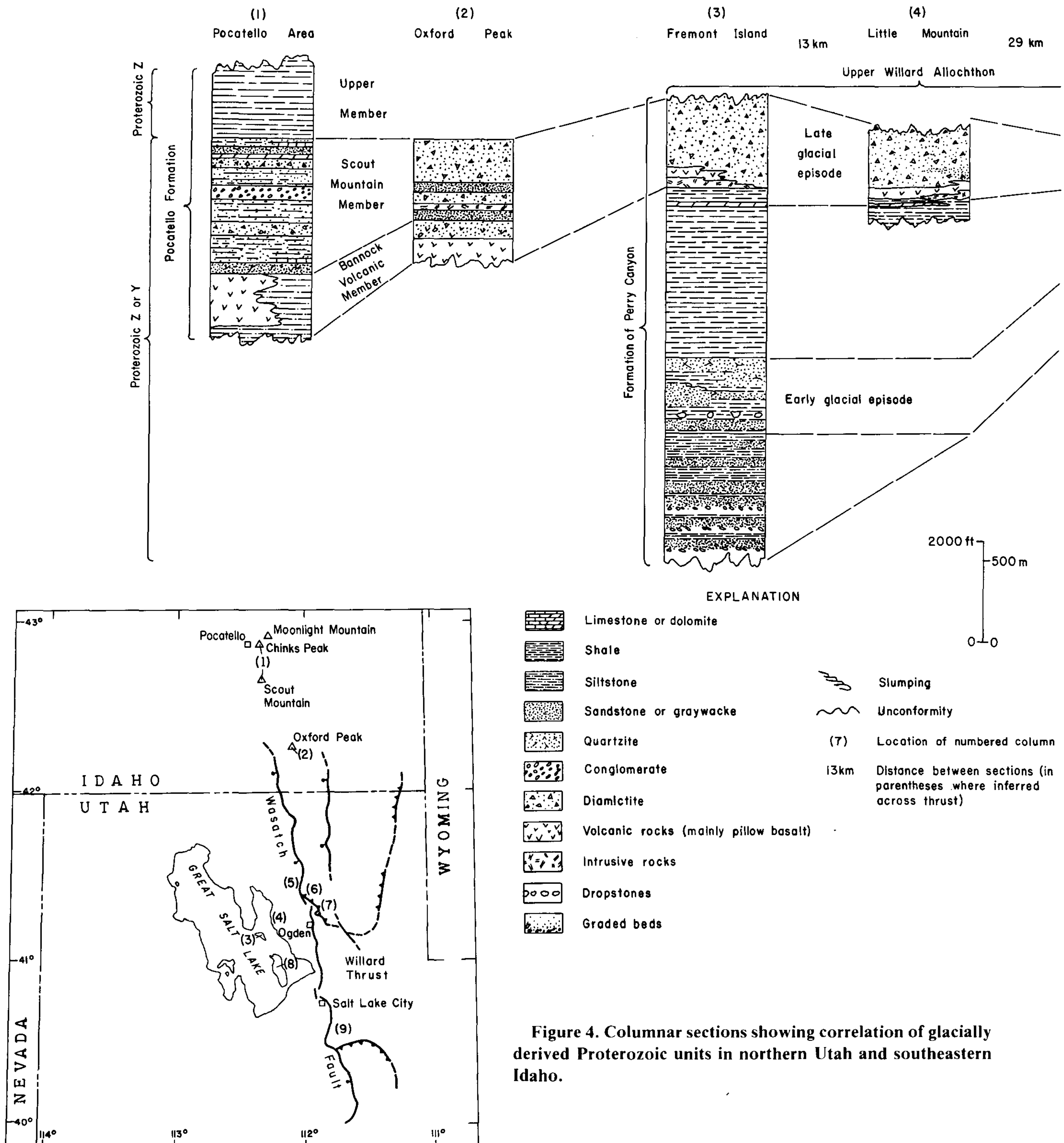


Figure 4. Columnar sections showing correlation of glacially derived Proterozoic units in northern Utah and southeastern Idaho.

thinner and less abundant, and the entire unit intertongues laterally first with blue-black, slaty, pyritic mudstone and this, in turn, with olive-drab siltstone and tan-weathering, coarse graywacke. The absence of free carbon, which results in "normal" colors, suggests that the graywacke and siltstone accumulated in an environment less strongly reducing than that represented by the diamictite, as well as one farther from the influence of glaciers.

Over much of the area, rocks of the lower thrust plate are slaty or phyllitic and show distinct foliation oblique to bedding. Chloritoid and chloritoid are widespread and, particularly along the ridge between Willard Peak and Brigham City, are large enough to form "spotted" phyllites. In a few places near Willard Canyon (Fig. 2), the graywacke and sandstones show incipient development of metamorphic biotite, but this is recognizable only in thin section.

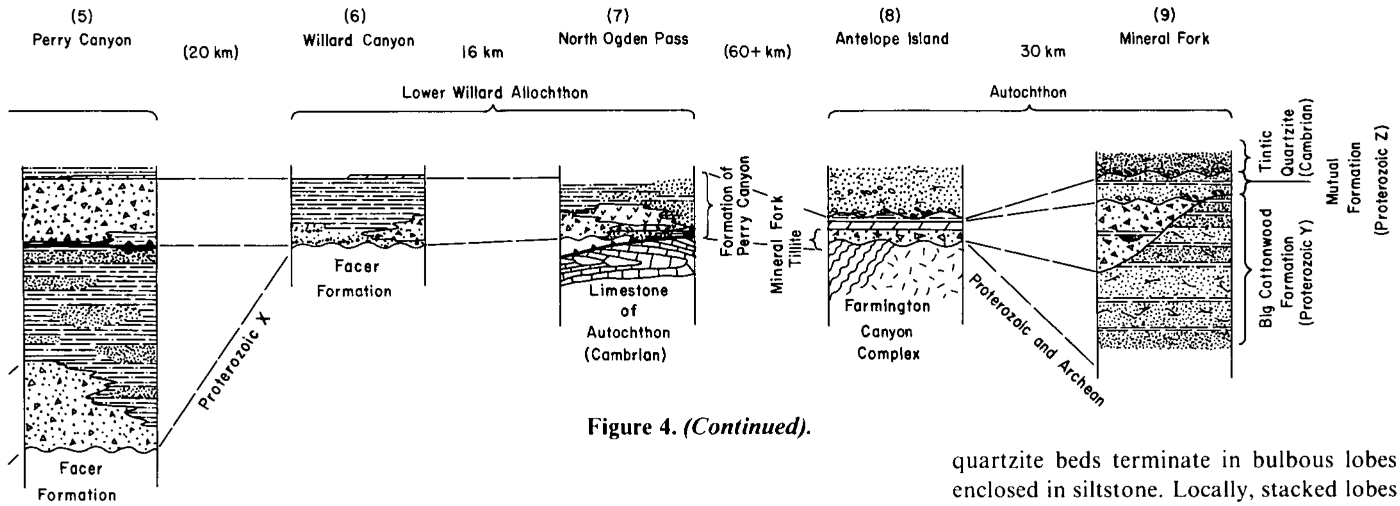


Figure 4. *(Continued).*

Fremont Island

At the west face of the Wasatch Range (Figs. 1 and 2), the Willard thrust is cut by the Wasatch fault and displaced southward, passing beneath the Great Salt Lake between Antelope and Fremont Islands (Crittenden, 1972). As a consequence, the westward extension of the allochthonous sequence exposed in Perry and Willard Canyons crops out on Little Mountain, Fremont Island, and nearby on the south end of Promontory Point. As noted above, the diamictite exposed only 10 km to the south on Antelope Island is structurally below the Willard thrust and therefore was deposited many tens of kilometres to the east of that in the allochthon.

The most distinctive unit of the rocks exposed on Fremont Island is a 600-m-thick section of diamictite that extends across the northern ridge of the island and forms its highest point (Eardley and Hatch, 1940b; Condie, 1967; Blick, 1979). These and underlying rocks that occupy the remainder of the island are here assigned to the formation of Perry Canyon (Fig. 4, column 3). Near the western point of the island, the diamictite intertongues with basalt flows and peperite and is cut by several shallow sills. Although relict textures are well preserved, the igneous rocks are altered to greenstone that consists mainly of chlorite, sericite, epidote, actinolite, albite, calcite, and iron oxides. These rocks have been searched repeatedly for material fresh enough to use for isotopic dating, but without success. This diamictite is characterized by large clasts of gneissic granite (as much as 5 m), many of which have been freed from the less-resistant matrix by wave action and now form storm beaches along the northern shore of the island. Quartzite, including a distinctive green chromian variety, is the next most abundant clast, and metamorphic and volcanic rocks and limestone form the remainder. Although the giant clasts are impressive and locally form closely packed masses, they are rather widely dispersed in a dark pyritic matrix in which the clasts are predominantly in the range of 1 to 5 cm in diameter. This body of diamictite is generally massive at the base but in the upper part is distinctly bedded on a scale of several metres, and it contains lenses and irregular beds of slate, siltstone, and, locally, tuff. Several examples of dropstones and flattened till clots were observed, although fine details of sedimentary structure commonly are obscured by the marked foliation. This unit appears to be the same as that exposed on Little Mountain to the east (Christie-Blick, 1980b).

Beneath the prominent diamictite, and forming the saddle across the center of the island, is a unit composed predominantly of black slate that is strongly foliated and locally knotted and spotted by chloritoid and pyrite. A section approximately normal to the strike indicates that the maximum possible thickness is on the order of 1,000 m, but because the entire unit is internally folded, the actual thickness is uncertain. Near the top of the slate unit, there is several tens of metres of distinctive clastic dolomite and dolomitic sandstone.

Below the slate, there is a section of intercalated dirty quartzite, siltstone, and locally, greenstone. Part of it is thrown into large-scale folds that are a prominent feature of the southwest face of the island. Evidence of slumping and wet-sediment deformation is also seen in several places where irregular quartzite beds terminate in bulbous lobes enclosed in siltstone. Locally, stacked lobes suggest that successive bodies of sand came to rest at about the same place. This lower part of the unit contains sparse dropstones, consisting of rounded quartzite clasts from 0.5 to 2.5 m in diameter, associated with sand clots, and enclosed in thin-bedded siltstone (Figs. 5 and 6). Although the clasts and clots are limited in number and distribution, they provide unequivocal evidence of an episode of glacial transport and rafting appreciably older than that recorded by the massive diamictite exposed across the northern part of the island.

The lowest units of the formation of Perry Canyon exposed on Fremont Island consist of graded beds of coarse grit from 1 to 5 m thick, intercalated with siltstone and mudstone (Fig. 4, column 3). Clast size in graded beds commonly ranges from coarse granules and small pebbles at the base to coarse sand (1 mm) at the top. All beds are strongly folded and locally overturned (Fig. 7). A few cross-beds in this unit, when restored for tectonic tilt, suggest northwestward paleocurrents. In view of the deformation, however, this direction must be regarded as tentative.

POCATELLO AREA, IDAHO

Diamictite in the Scout Mountain Member of the Pocatello Formation near Pocatello has been attributed to glaciomarine sedimentation by all who have studied it (Ludlum, 1942; Crittenden and others, 1971; Trimble, 1976; Link and others, 1980; Link, 1981, 1982, 1983), on the basis of the presence of striated clasts in diamictite and correlation with glacial rocks in northern Utah. Recent mapping (Link and others, 1980) has shown, however, that part of the type section of the Pocatello Formation is structurally overturned, and that pyritic

Figure 5. Dropstones, the largest 2.5 m long, enclosed in dark brown-weathering siltstone in lower glacial unit, Fremont Island.

shale originally designated the "lower member" by Trimble (1976) is actually part of the upper member. A revised stratigraphic column of the Pocatello Formation is shown in Figure 4 (column 1).

The Pocatello Formation is exposed along much of the length of the Bannock Range from Moonlight Mountain to south of Oxford Peak (Fig. 4). The most complete section through the diamictite-bearing Scout Mountain Member occurs on both sides of the north-trending ridge of Scout Mountain 20 km south of Pocatello. The lowest exposed unit is a greenish, vitreous, sparsely pebbly quartzite which, if unfaulted, is 280 m thick. A poorly exposed and lenticular interval of greenstone, breccia, and porphyritic lava assigned to the Bannock Volcanic Member overlies the quartzite. These volcanic rocks are 110 m thick on Scout Mountain and thicken northward. Next above them, there is a 70-m interval of brown, thin-bedded sandstone, followed by the lower diamictite, which is about 100 m thick. The lower diamictite is crudely bedded, and grades upward into inversely to normally bedded sandstone. Five pebble counts in this diamictite average 43% quartzite, 52% siltstone, and 5% volcanic rocks. A 150-m coarsening-upward sequence of siltstone and graded sandstone overlies the lower diamictite, culminating in 50 to 100 m of cobble and boulder conglomerate that is intercalated with trough-cross-bedded medium- and coarse-grained sandstone. One pebble count in the conglomerate indicated 85% quartzite, 11% granitic rocks, 2% siltstone, and 2% volcanic clasts.

Quartzite and brown siltstone overlie the conglomerate and are in turn overlain by the upper diamictite, which is more than 150 m thick and contains clasts as much as 1.5 m in diameter. Striated clasts are present, although rare. Two pebble counts in the upper diamictite on Scout Mountain average 28% quartzite, 51% basement rock (granitic and gneissic rocks), 20% siltstone, and 1% volcanic rock. A 1-m-thick laterally persistent bed of laminated, pink silty dolomite caps the upper diamictite. The section above the upper diamictite is well exposed in structurally overturned beds north of Portneuf Narrows (Fig. 1), where a thinning- and fining-upward transgressive clastic and limestone sequence overlies the dolomite. The upper member of the Pocatello Formation is a monotonous dark gray, silver-weathering laminated and locally pyritic shale or argillite that is similar to the Kelley Canyon Formation of Utah and to postglacial shales (Twitya Formation) at a similar stratigraphic level in the Northwest Territories, Canada (Christie-Blick and others, 1980).

On the slopes west of Chinks Peak (Fig. 4) diamictite of the Scout Mountain Member is interbedded with and contains clasts of the Bannock Volcanic Member, which here reaches a thickness of 400 m. Both diamictite and cobble conglomerate are present southeast of Chinks Peak and in an isolated northern exposure on Moonlight Mountain. The diamictite is thicker at these

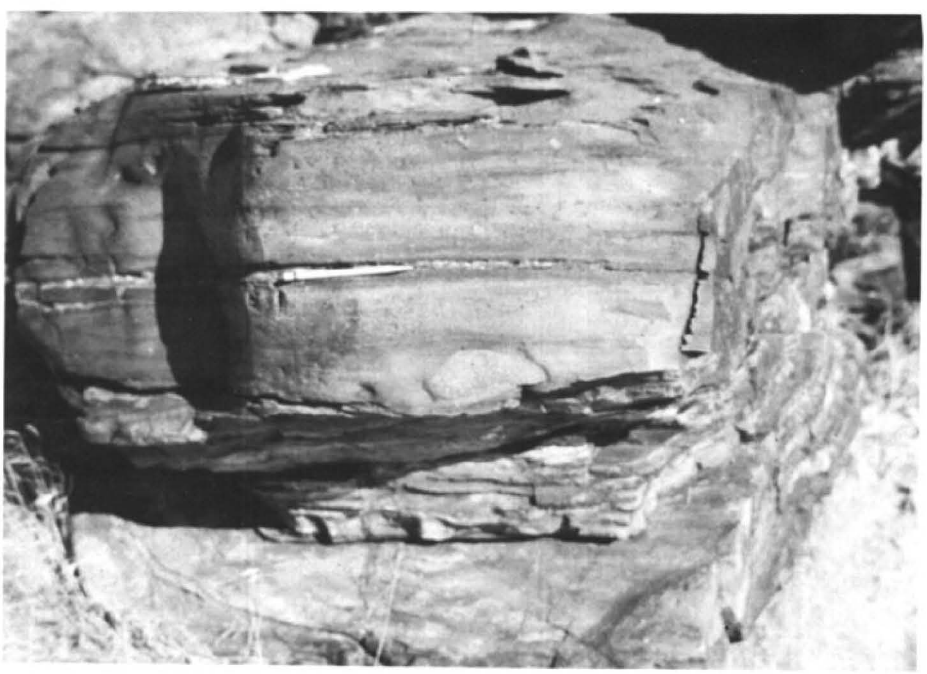


Figure 6. Sand clot 12 cm long, inferred to have been dropped into thinly laminated siltstone, lower glacial unit, Fremont Island.

localities than on Scout Mountain, but structural complications preclude accurate measurement of the section.

The Scout Mountain Member near Pocatello consists of a variety of sedimentary facies (Link, 1982, 1983). The thickest diamictite is exposed in the north on Chinks Peak and Moonlight Mountain (Fig. 4), where it is interpreted as lodgement tillite or glaciomarine diamictite, in part deposited by sediment gravity flow. South of Portneuf Narrows, slope and basinal facies are present, including strata interpreted as turbidites and "contourites," mass-flow channelized conglomerate, and redeposited diamictite. The southernmost exposures at Garden Creek Gap (Fig. 1) consist predominantly of shallow-marine and nearshore deposits (Thompson and Link, 1981; Thompson, 1982).

OXFORD PEAK, IDAHO

At least 450 m of diamictite, graywacke, and siltstone of the Scout Mountain Member of the Pocatello Formation occurs on Oxford Peak (Fig. 1) and in isolated hills in Cache Valley to the east. These clastic rocks overlie and are intruded by mafic subaqueous volcanic rocks of the Bannock Volcanic Member of the Pocatello Formation that include pillow lava, agglomerate, volcaniclastic siltstone, and diabase sills. All are metamorphosed to greenschist facies.

The Scout Mountain Member is laterally variable, in both thickness and lithology. In its southernmost exposures at Five Mile Canyon, some 15 km southeast of Oxford Peak (Fig. 1), it contains 250 m of massive cobble-bearing diamictite. About 1 to 2 km south of Oxford Peak, it consists of 200 m of medium- to thick-bedded cobble diamictite, silt-chip diamictite, siltstone, and fine graywacke. These beds interfinger to the south with green volcaniclastic sandstone. Near Oxford Peak, a 40-m-thick sill of metadiabase intrudes silt-chip diamictite and extends about 10 km to the southeast. A generalized section is shown in column 2 of Figure 4.

Thirteen pebble counts in diamictite on Oxford Peak averaged 28% white, gray, and brown quartzite; 10% granite and vein quartz; 39% black and light gray siltstone; 10% medium and coarse sandstone; and 13% volcanic clasts, including white felsite and dark porphyry.

The massive diamictite at Five Mile Canyon is inferred to have been deposited as lodgement till near the grounding line of an ice sheet that reached the sea. The heterogeneous beds along the ridge south of Oxford Peak are interpreted as diamicton deposited by sediment gravity flow and sediment derived by the winnowing of diamicton, intercalated with volcanic sand and pillow lava. A common and distinctive lithology is silt-chip diamictite with a small component of quartzite or granitic clasts. The silt fragments are thought to have been incorporated during redeposition of diamicton on a submarine slope.

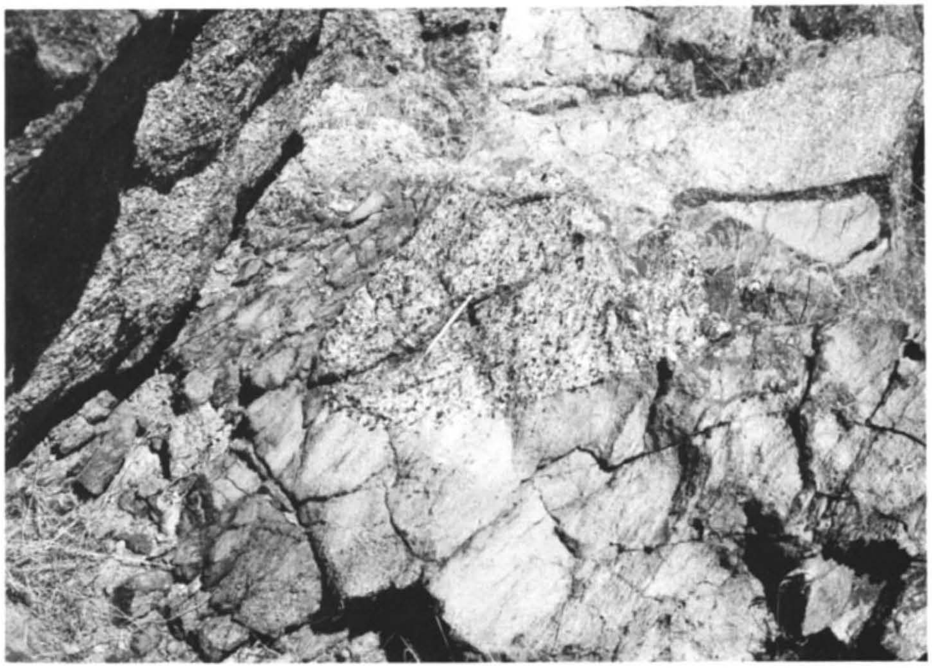


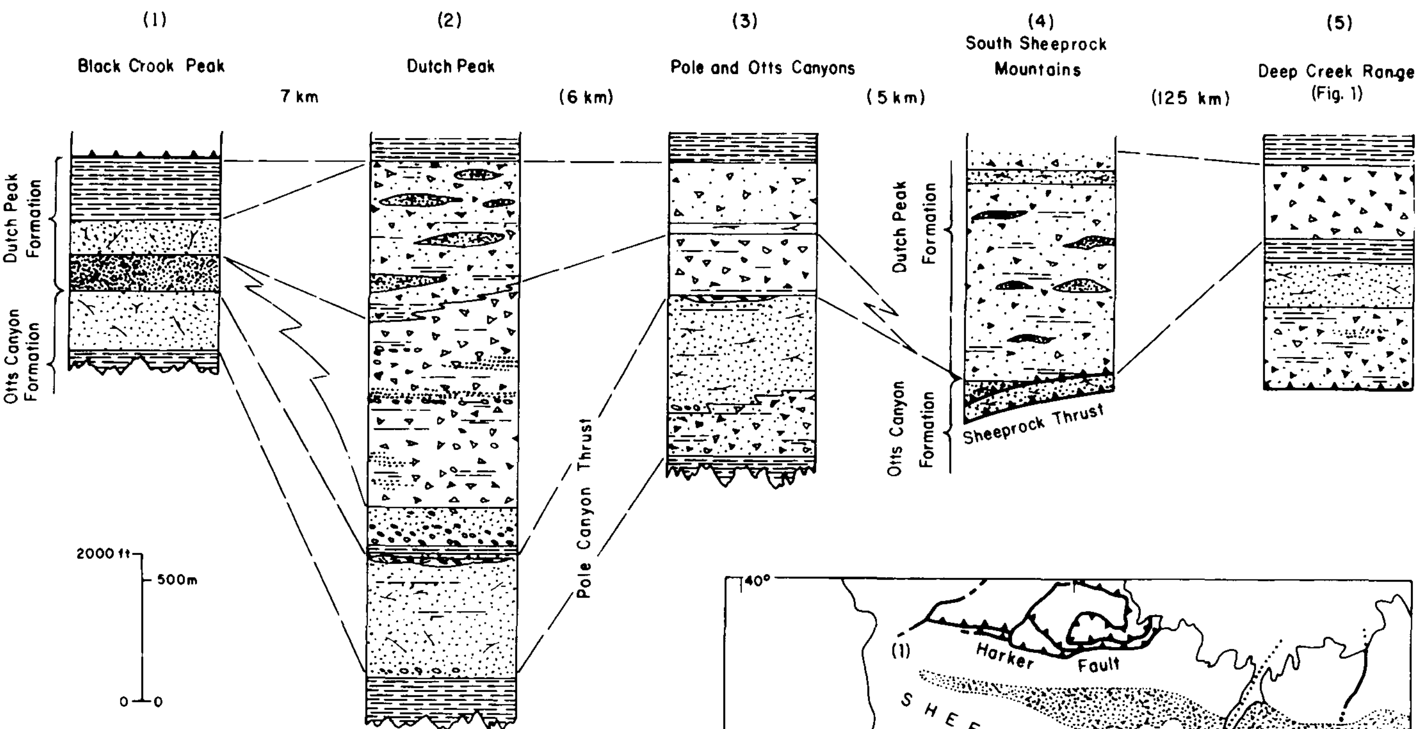
Figure 7. Folded coarse-grained graded beds below lowest glacial unit, Fremont Island. Pencil is 12 cm long.

SHEEPROCK MOUNTAINS

A thick sequence of little-metamorphosed, allochthonous Proterozoic rocks, including diamictite, occurs in the Sheeprock Mountains, about 135 km south of Fremont Island and 60 km west of the Wasatch Mountains (Figs. 1 and 8). Several major thrusts separate these rocks from those immediately above the Willard thrust, but it seems likely that the rocks of Sheeprock Mountains were transported about the same distance eastward with respect to the autochthon as was Fremont Island (Blick, 1979).

Proterozoic rocks were early recognized in the Sheeprock Mountains by Loughlin (1920) and the over-all stratigraphy was established in mapping by Cohenour (1959) and others in the 1950s. Subsequent mapping by Morris and Kopf (1970a, 1970b) and Christie-Blick (1982b) has led to further stratigraphic subdivision and recognition of individual formations of the Huntsville sequence of Crittenden and others (1971). Diamictite occurs in two stratigraphic units, the lower one in the middle of the Otts Canyon Formation and the upper one in the overlying Dutch Peak Formation (Dutch Peak "tillite" of Cohenour, 1959). The lower diamictite is thin and lenticular and is separated from the thick upper body by several hundred metres of quartzite.

The lower diamictite unit is best exposed in the vicinity of Otts Canyon (Fig. 8, column 3), where it is from 0 to 500 m thick and consists of interbedded diamictite, graywacke, grit, quartzite, and slate. Underlying rocks consist of several hundred metres of gray and silver-gray, banded slate and phyllite, lithologically similar to pelitic rocks interbedded with the diamictite, and suggestive of a conformable and perhaps interfingering contact. The diamictite occurs in intervals from a few metres to several tens of metres thick and is generally in sharp contact with adjacent slate and quartzite. The diamictite matrix is gray, green, or black, and phyllitic; texturally, it is fairly homogeneous to heterogeneous with silty and gritty wisps . The concentration of



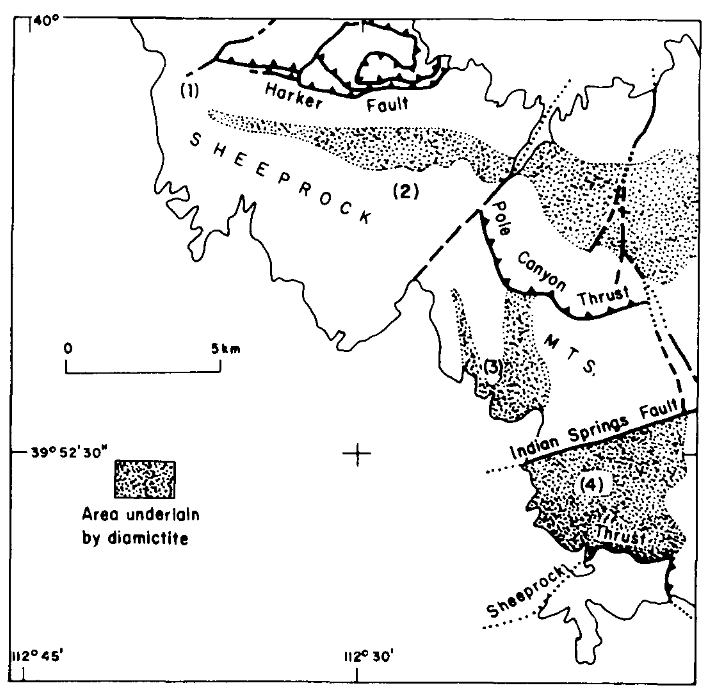


Figure 8. Columnar sections showing correlation of glacially derived Proterozoic units in the Sheeprock Mountains and Deep Creek Range, central and western Utah. Lithologic symbols as in Figure 4. Index map is for Sheeprock Mountains only. Deep Creek Range is located in area shown in Figure 1.

clasts is variable; typically these consist of about 95% quartzite, with subordinate quartz, dolomite, granite, and intraformational siltstone. Dolomite is locally abundant, however. Clasts are predominantly 2 to 5 cm in diameter, but a few are as large as boulders. Beneath the Pole Canyon thrust, in the vicinity of Otts Canyon, the lower diamictite unit thins abruptly toward the north. Above the Pole Canyon thrust, it is less than 70 m thick and pinches out toward the northwest. The lower diamictite is not exposed south of the Indian Springs tear fault (Fig. 8, column 4), probably because the exposures do not extend far enough downsection at this locality (Blick, 1979).

Between the diamictite units, and constituting the upper member of the Otts Canyon Formation (Fig. 8, column 3), there are several hundred to more than 1,000 m of gray quartzite, with subordinate interbeds of conglomerate and grit (especially near the base), and of graywacke, slate, siltstone, and shale (especially near the top). The upper part is intruded by diabase sills. The quartzite is typically fine to medium grained, with moderately to well-rounded and sorted grains, and is locally vitreous. It is thin to thick bedded, commonly laminated, and rarely cross-bedded. Rare, angular fragments of black diamictite in conglomerate near the base confirm the stratigraphic sequence and indicate that the diamictite of the early glacial episode was consolidated before the overlying beds were deposited. The quartzite interfingers with the underlying diamictite unit and is markedly thinner toward the south in rocks beneath the Pole Canyon thrust. Near the divide between Otts and Pole Canyons (close to location 3 in Fig. 8), however, the contact between the quartzite and lower diamictite units is probably faulted, and the exposed thickness of quartzite may be less than its true stratigraphic thickness. Above the Pole Canyon thrust, the quartzite unit thins toward the northwest and is less than 250 m thick near Black Crook Peak.

The upper diamictite unit, the Dutch Peak Formation, consists of as much as 1,750 m of moderately bedded to well-bedded diamictite, conglomerate, graywacke, grit, sandstone, quartzite, siltstone, and shale. The contact with the underlying quartzite is transitional but probably not interfingering, because a bed of distinctive metallic blue-gray quartzite, associated with dropstones near the base of the Dutch Peak, persists along the northern flank of the Sheeprock Mountains for several kilometres at the same stratigraphic level. Angular diabase fragments, locally included in conglomerate above this marker bed, were probably derived from the underlying diabase sills or compositionally similar rocks at about the same stratigraphic level. The diabase fragments thus attest to local erosion (channels?) and to penecontemporaneous mafic volcanism as in other parts of the allochthon. Poorly sorted rocks of the Dutch Peak Formation are typically olive green to gray, chloritic, moderately phyllitic, and mineralogically, chemically, and texturally heterogeneous at microscopic to megascopic scales (Blick, 1979). Major clast types are dolomite, granite, and quartzite (including the rare but distinctive chromian variety). Subordinate types include a variety of metamorphic, igneous, and sedimentary rocks. In the Sheeprock area, 29 clast counts indicate pronounced lateral and stratigraphic variations in the proportions of clast types over short distances, but some gross generalizations can be made. Dolomite clasts are most abundant near the base and toward the east, quartzite clasts are most common near the top and south of the Indian Springs fault, and igneous clasts are particularly common in the northwest. Clasts are predominantly 2 to 5 cm in diameter, although rarely as large as 3 m (one granite gneiss and one chromian quartzite). Rocks above the Pole Canyon thrust are coarsest near the base. The size of largest clasts decreases stratigraphically upward and toward the south. Diamictite and conglomerate of the Dutch Peak Formation intertongue with lenticular, mature quartzite in the upper part and in southern exposures. In addition, thick, poorly sorted rocks pass laterally into rather thinner quartzite toward the northwest in the vicinity of Black Crook Peak (Fig. 8, columns 1 and 2).

The thinning of both diamictite units and the intervening quartzite toward the northwest and the facies change of the upper diamictite to quartzite in the same direction suggest deposition near the edge of a locally subsiding marine embayment. The diamictite is interpreted to be mainly a basinal deposit, and the quartzite to represent better-sorted shelf sand that prograded into the adjoining basin. The bedding characteristics and textural heterogeneity of the diamictite and the association at many stratigraphic levels of diamictite with laminites containing dropstones suggest that the diamictite was deposited by sediment gravity flow and from floating ice. Paleocurrents estimated from the few available structures in the upper diamictite and the medial quartzite are directed toward the southwest, perhaps approximately parallel to the edge of the inferred basin. The Sheeprock basin was characterized by generally oxidizing conditions and differs in this respect from the Perry Canyon and Mineral Fork basins, where conditions were reducing.

DEEP CREEK RANGE

Metadiamictite, presumably in the western part of the allochthon, crops out about 125 km west of the Sheeprock Mountains in the southern part of the Deep Creek Range in two units separated by about 200 m of quartzite (Misch and Hazzard, 1962). Bick (1966) reinterpreted the outcrop pattern in terms of a single diamictite unit repeated by a tight syncline, with a core of quartzite. Facing directions are difficult to determine from available sedimentary structures because the rocks are in the garnet and staurolite grades of regional metamorphism (Nelson, 1969) and have a strong schistose fabric. Nevertheless, important lithologic differences between the diamictite units seem to corroborate the original interpretation of Misch and Hazzard (1962). The lower unit consists of sparsely pebbly schist, which appears to be less sandy and gritty and better bedded than the generally more crowded pebbly diamictite of the upper unit (Christie-Blick, 1982b). Clasts, mostly the size of pebbles but rarely as large as 1 m, are mainly "granite" (gneissose and directionless varieties; about 70%) and quartzite (about 25%). Subordinate types are other metamorphic, igneous, and sedimentary rocks, including marble. The interpretation of the diamictite units as glaciomarine (Misch and Hazzard, 1962) is tentatively reaffirmed on the basis of well-developed bedding and the presence of a possible dropstone near the top of the upper unit.

CORRELATION OF GLACIAL DEPOSITS WITHIN UTAH

Until recently, stratigraphic data have been inadequate to permit detailed correlation of glacial deposits between any of these sections of Proterozoic Z age. Mapping described above, however, now suggests that the glacial episodes recorded by thick bodies of diamictite or by the presence of dropstones can be correlated confidently at least within northern and central Utah. The correlations currently proposed (Figs. 4 and 8) are based on stratigraphic position and on the presence of intercalated or closely associated volcanic rocks and, locally, of carbonate rocks. They are strongly supported also by unit-by-unit correlation of the overlying non-glacial formations that constitute the remainder of the Huntsville sequence (Crittenden and others, 1971; Christie-Blick, 1982b).

Correlation within the Willard allochthon is comparatively straightforward. The presence of intercalated pillow basalt and the distinctive underlying gray clastic dolomite permits a direct tie from the thick diamictite on Fremont Island to that exposed on Little Mountain (Christie-Blick, 1980b). From there to exposures in the Wasatch Range, the tie is less explicit, but in the upper allochthon at Perry Canyon, the upper diamictite also contains thick lenses of pillow basalt in exposures 1.5 to 2.5 km east of the canyon mouth, whereas igneous rocks are thin or absent in the lower diamictite. Correlation of the upper diamictite with diamictite in the lower Willard allochthon also is supported by pillow basalts and mafic to intermediate intrusives present in that unit, particularly near North Ogden Pass (Fig. 4, column 7). The formation of Perry Canyon as a whole thins eastward, and the upper diamictite horizon rests directly on the Willard thrust in the southeasternmost exposures in Ogden Canyon. This is to be expected from the observation that over a larger area the Willard thrust rises stratigraphically eastward.

Correlation of the glacial units within the allochthon from Perry Canyon and Fremont Island to the Sheeprock Mountains, 135 km to the south (Fig. 8), is much more difficult, even though all the major units of the Huntsville sequence of Crittenden and others (1971) except the Maple Canyon Formation can be recognized in both areas (Christie-Blick, 1982b). This difficulty is reflected in the use of different names (the Otts Canyon and Dutch Peak Formations) for the glacial units. The Kelley Canyon Formation, which overlies the Maple Canyon in the northern Wasatch Range, rests directly on the Dutch Peak in the Sheeprock Mountains. The greatest difference between the two sequences is that quartzite intervenes between the diamictite units in the Sheeprock Mountains, whereas graywacke and siltstone occupy this interval in

Perry Canyon and slate does so on Fremont Island. In addition, the diamictites in the two areas are petrographically quite distinct. On the other hand, the lower glacial unit in both areas overlies other "basinal" deposits, and the upper unit in both is associated with mafic volcanic or intrusive rocks. These similarities suggest that the glacial units in the two areas may be correlative in spite of the differences between them.

Units of the Huntsville sequence have not been recognized in the Deep Creek Range, probably for the most part as a result of lateral facies changes, but also as a result of metamorphism and structural complications. Regardless of this problem, the tie from the Sheeprock Mountains westward to the Deep Creek Range is perhaps more secure than correlations to the north and east. In the Deep Creek area, diamictite units are separated by mature quartzite and overlain by schist that may be equivalent to the Kelley Canyon Formation (Fig. 8, column 5). Although the rocks differ in detail from those in the Sheeprock Mountains, the over-all sequence is very similar.

Correlation from the allochthon to the autochthon within Utah also remains problematic; the mafic igneous rocks, preferentially associated with the upper diamictite within the allochthon,are entirely absent in the autochthon. The capping layer of pinkish-tan, laminated dolomite in the autochthonous or parautochthonous section of diamictite on Antelope Island (Fig. 4, column 8), however, closely resembles the laminated dolomites that locally overlie the diamictites of Perry Canyon and the conglomerate beds of the Maple Canyon Formation. The closest exposure of such laminated dolomite in the allochthon is at the south tip of Promontory Point, where it appears to be associated with coarse, dark-gray, granule-bearing quartzite assigned to the Maple Canyon Formation that may be a facies equivalent of diamictite. Unfortunately, the detailed stratigraphic relations of these units to the rocks on Fremont Island are obscured by the intervening arm of Great Salt Lake. The best current inference is that the diamictite on Antelope Island represents a thinner shoreward equivalent of the upper diamictite of the Fremont Island, Little Mountain, and Perry Canyon sections. Such a correlation would be consistent with a general west-dipping paleoslope beneath the diamictite, combined with onlap of basin deposits toward the east. Some support is offered also by clast lithology; gneissic granite predominates in both allochthonous and autochthonous deposits, and one of us (Crittenden) has observed that clasts of the distinctive green chromian quartzites are a characteristic although minor constituent in both sections. Palinspastic restoration of the Willard thrust, however, indicates that the sections now exposed in Perry and Willard Canyons were originally deposited somewhere northwest of the present site of Antelope Island. Therefore, because the upper diamictite thins and becomes lenticular southeastward in the allochthon, the diamictites on Antelope Island may never have been directly continuous with those in the allochthon.

The Mineral Fork Tillite in the Cottonwood area (Fig. 4, column 9) lies nearly 30 km farther southeast and contains still fewer clues as to possible correlation. Neither basic volcanic rocks nor persistent bedded carbonate rocks are present. Moreover, clast composition is dominated by quartzite and dolomite, suggesting derivation from a sedimentary terrane, rather than one dominated by gneiss and granite. Although carbonate rocks are entirely absent from the underlying Big Cottonwood Formation and from the correlative Uinta Mountain Group that lies to the east along the trend of the subglacial valleys in which the Mineral Fork was deposited, it seems most likely that the abundant carbonate clasts in the Mineral Fork, which include oolitic and stromatolitic dolomites, were derived from a stratigraphically higher part of one or the other of those sequences of Proterozoic Y age. Such carbonate rocks are inferred to have been removed from the Uinta Mountain area by erosion during late Proterozoic glaciation or prior to deposition of the Cambrian and Mississippian rocks that now overlie the rocks of the Uinta Mountain Group in the western part of the range.

CORRELATION FROM NORTHERN UTAH TO SOUTHEASTERN IDAHO

Intercalated diamictite and mafic volcanic and shallow intrusive rocks assigned to the Pocatello Formation and exposed near Oxford Peak (Fig. 4, column 2) are correlated with the upper diamictite of the formation of Perry Canyon in the Willard allochthon. The outcrops of pillow basalt, volcaniclastic siltstone, and massive diamictite on Oxford Peak and near Twin Lakes Reservoir in Cache Valley to the east are strikingly similar to the exposures on Little Mountain near Ogden. Near Pocatello (Fig. 4, column 1), two thin units of diamictite are present in the Pocatello Formation, but they occupy a very restricted stratigraphic range and locally intertongue with the volcanics. Where both diamictites are present, they overlie the main mass of volcanic rocks. Therefore, the diamictites near Pocatello are also thought to correlate with only the younger of the two episodes of glaciation represented in the formation of Perry Canyon in Utah.

REGIONAL CORRELATION

The general equivalence of glacial rocks of late Proterozoic age throughout the Cordillera is now widely accepted (Stewart, 1972; Christie-Blick and others, 1980), but in most areas, only one episode of glaciation has been recognized. In view of the evidence presented here for two such episodes in parts of Utah, a similar possibility needs to be examined for other areas.

In the Death Valley area of California, thick bodies of diamictite in the upper Proterozoic Kingston Peak Formation were recognized long ago by Hazzard (1937), and it seems likely that these units are generally correlative with the glacial deposits in Utah and Idaho described above. Nevertheless, although these bodies locally reach nearly 3,000 m in thickness, and in places record fluctuations of an ice margin (Miller, 1982), detailed stratigraphic and paleogeographic studies completed to date do not yield evidence for more than one widespread episode of glaciation (Wright and others, 1976; Miller and others, 1981).

The southeastern part of the Mackenzie Mountains of the Northwest Territories, Canada, is the only other area where upper Proterozoic strata may record two episodes of glaciation. Eisbacher (1978) showed that locally, north of the Redstone River, 15 to 20 m of diamictite occurs at the base of the Sayunei Formation, and it is separated from the massive diamictites of the overlying Shezal Formation by about 250 m of maroon siltstone (see section 7 of his Fig. 10). This thickness is approximately comparable with that separating the diamictite units in Utah, suggesting that two pulses of glaciation may be represented here also. However, both Young (1976) and G. H. Eisbacher (1980, written commun.) noted that the siltstones grade laterally into turbidites containing abundant outsized clasts believed to be dropstones. Thus, over a larger area, glaciation may have been relatively continuous, and there is no firm basis for detailed correlation with the two glacial intervals described in Utah.

PALEOGEOGRAPHIC SETTING AND SEDIMENTATION RATE

The data now available suggest that early in Proterozoic Z time, the present site of central Utah and adjoining parts of western North America were situated within a continent that was undergoing regional extension and glaciation. Glacial deposits occur intermittently throughout the Cordillera from Death Valley to Alaska (Stewart, 1972; Christie-Blick and others, 1980). Tongues of ice periodically extended into the adjoining sea and dropped their entrained debris on the edge of the continental platform, and in graben where it mingled with normal marine sediments. Glacial deposits that may be similar in age are present in the southern Appalachian Mountains (Schwab, 1976; Rankin and others, 1969), but because no traces of late Proterozoic glaciation are preserved in the central part of North America, the possible extent of ice sheets across the craton cannot be determined.

The best-documented lithologic analogues for these Proterozoic deposits are from the Antarctic (Barrett, 1975; Anderson and others, 1981) and the Miocene through Holocene deposits of the Yakataga Formation off the coast of Alaska (Plafker and Addicott, 1976). Although the latter are inferred to be the product of alpine glaciation along a rugged coastline rather than a low-lying one, the sediments bear a close resemblance to those of the formation of Perry Canyon and the Mineral Fork Tillite. Diamictite is abundant in all sections of the Yakataga, locally making up more than 60% of the formation. Siltstone and mudstone vary greatly in abundance, as in Utah. However, sandstone and conglomerate (outwash?) appear to be more common in the Yakataga than in the Perry Canyon, and in this respect, the Yakataga more closely resembles the Mineral Fork Tillite.

Although sediment accumulation rates are notoriously variable (Sadler, 1981), the time interval that separated the late Proterozoic glacial episodes proposed in this paper can be evaluated if appropriate rates are assumed. Barrett (1975) estimated an average accumulation rate of 40 m/m.y. for several hundred metres of silty claystone with sparse pebbles in the Ross Sea. Plafker and Addicott (1976) calculated a considerably higher sedimentation rate of about 1,000 m/m.y. for nearly 1,200 m of Pleistocene diamictite and conglomerate on the outer continental shelf of the Gulf of Alaska. If these rates are applicable to the formation of Perry Canyon, the glacial episodes may have been separated by as little as 0.5 m.y. or as much as 25 m.y.

ACKNOWLEDGMENTS

The data reported here are by-products of geologic mapping by the U.S. Geological Survey (Crittenden and associates in Utah and S. S. Oriel and associates in Idaho) beginning as early as 1950 and continuing intermittently through 1981. This research included detailed mapping in the central and northern Wasatch Range and reconnaissance on Antelope and Fremont Islands in the Great Salt Lake. Recent geologic mapping and sedimentologic studies by Christie-Blick in the Wasatch Range, on Fremont Island, at Little Mountain, and in the Sheeprock Mountains, and by Link in southeastern Idaho were supported by National Science Foundation Grants ATM 74-24201, EAR 77-06008, and EAR 78-15194 to J. C. Crowell. Responsibility for this report is shared jointly, except for the parts dealing with the Sheeprock Mountains and Deep Creek Range, Utah, prepared by Christie-Blick, and the section on southeastern Idaho, prepared by Link. We are indebted to H. Doelling, J.M.G. Miller, H. T. Morris, and L. A. Woodward for helpful reviews of the manuscript.

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Manuscript Received by the Society November 9, 1981
Revised Manuscript Received May 25, 1982
Manuscript Accepted June 9, 1982