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KENT C. CONDIE *Dept. of Earth Sciences, Washington University, St. Louis, Missouri*

# Petrology of the Late Precambrian Tillite(?) Association in Northern Utah

**Abstract:** The late Precambrian subgraywackes of northern Utah and adjacent areas are poorly sorted and commonly graded, and contain bent and twisted, reworked subgraywacke fragments. Such features suggest a turbidity current origin. The laminated subgraywackes differ strikingly from Pleistocene varves of the same thickness in that they contain a considerably larger amount of coarse particles and that their carbonaceous matter is not concentrated at the top of each bed. The presence of bent and twisted subgraywacke fragments and contorted beds in the conglomeratic subgraywackes (tillites?) and the common lateral gradation of these rocks into subgraywackes suggests that they are of subaqueous mudflow origin.

Although textural and structural evidence in the subgraywackes and conglomeratic subgraywackes suggests a turbidity current and subaqueous mudflow origin, the possibility of contemporary glaciation cannot be eliminated inasmuch as glacial sediments may be redeposited by these agents. The possibility of reworking, mixing, and redepositing of sediments derived from contemporary (or extinct) glaciers with sediments of nonglacial origin greatly complicates and temporarily renders insoluble the problem of determining the prevailing climatic conditions at the time these rocks were redeposited.

The subgraywackes and conglomeratic subgraywackes consist chiefly of rounded clastic quartz and minor feldspar grains together with rock fragments in a diagenetic or metamorphic matrix of predominantly mica, chlorite, and quartz. Although the preserved clastic components suggest plutonic- and quartzite-rich source areas, the nature and origin of the original matrix minerals is unknown. It is suggested that they were derived from erosion and redeposition of a weathered mantle on plutonic source areas.

Differences in mineralogy and major and trace element concentrations occur between unmetamorphosed and low-grade metamorphic subgraywacke and conglomeratic subgraywacke sections. Such differences appear to be related to a combination of variability in original sediment compositions and compositional changes accompanying progressive metamorphism. Widespread compositional uniformity of low-grade metamorphic plagioclase, chlorite, and to a lesser degree muscovite and paragonite, suggests that chemical equilibrium was approached in these phases during metamorphism.

CONTENTS



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## INTRODUCTION

The late Precambrian rocks of northern Utah, and vicinity are of particular interest in that they contain subgraywackes[1] and conglomeratic subgraywackes which have been interpreted as lithified glacial varves (varvites) and tills (tillites). Similar presumed glacial deposits occur on most of the continents in rocks of late(?) Precambrian age. Because some of these rocks, which are commonly interbedded with quartzite and carbonate rocks in thick geosynclinal sections, have been reported to possess textures and structures suggestive of subaqueous mudflow and turbidity current deposition, their presumed glacial origin and glacial epochs postulated therefrom are open to question.

This investigation reports field and laboratory data bearing on the origin, diagenesis, and metamorphism of the late Precambrian subgraywackes and conglomeratic subgraywackes and associated rocks in the Wasatch Range and Great Salt Lake area in northern Utah (Fig. 1). A few samples were also studied from other late Precambrian terranes in the northeastern Great Basin. In order to understand better the geologic setting, stratigraphy, and large-scale textures and structures, three typical areas in northern Utah were mapped and studied in detail. The locations of these areas, the Big Cottonwood Canyon area, Fremont Island in Great Salt Lake, and the southern Promontory Range, are shown in Figure 1.

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## FIELD STUDIES

### *The Big Cottonwood Canyon Area*

Unmetamorphosed late Precambrian rocks are exposed in several small areas in the central Wasatch Range. The most extensive and best-exposed section occurs in the Big Cottonwood Canyon area southeast of Salt Lake City. This area has been mapped and its geologic features discussed by Crittenden and others (1952). The late Precambrian rocks of this area (known as the Big Cottonwood Series) were briefly described by Eardley and Hatch (1940, p. 819–820) and by Crittenden and others (1952, p. 3–6). The author mapped in detail an area of about 50 km$^2$ which is shown in Figure 2. The late Precambrian rocks of this area occur on the southern limb of a large southeast-trending syncline in the central Wasatch Range. Section AA′ in Figure 2 is a cross section of part of this limb. The area mapped encompasses a late Precambrian section chiefly of quartzite with lesser amounts of subgraywacke and conglomeratic subgraywacke totaling about 4600 m thick. This sequence, which lies unconformably

[1] Quartzitic sandstones, siltstones, and associated phyllitic shales with quartz/feldspar ratios ≥ 9 and quartz + feldspar/mica + chlorite ratios ≤ 4 (after F. Pettijohn in Krumbein and Sloss, 1955, p. 130).

on an older Precambrian complex immediately southwest of the area mapped, reaches a maximum thickness of about 5500 m.

The subgraywackes and conglomeratic subgraywackes, which have been previously interpreted as glacial deposits (Hintze, 1913, p. 91–103; Blackwelder, 1932, p. 297; Crittenden and others, 1952, p. 4–6) are well exposed in the Big Cottonwood Canyon area. The conglomeratic subgraywacke, as mapped in Figure 2, is composed of elongate lenticular conglomeratic subgraywacke and minor conglomeratic quartzite units (from 1 to 7 m thick and many tens of meters in length) interbedded with subgraywacke and quartzite. The entire mass is broadly lenticular with a maximum exposed thickness of 1700 m; however, because the upper contact is an erosional unconformity, the original thickness is unknown. The interbedded subgraywacke could not be distinguished at the scale at which the mapping was done. However, some of the larger interbedded quartzite units are shown on the geologic map. Subgraywacke (and thin-bedded quartzite) without associated conglomeratic subgraywacke also occurs lower in the section and is shown as such on the geologic map. A measured section was made by Dott (1961, p. 1299) in the Mineral Fork area.

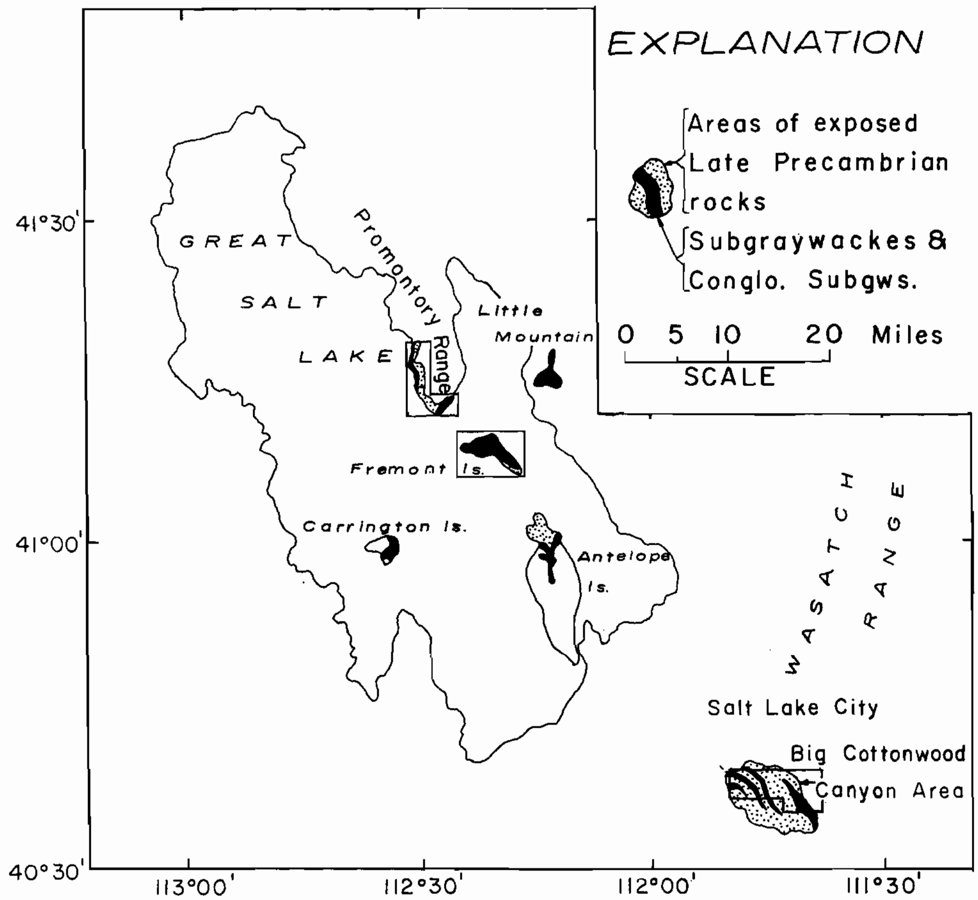



Figure 1. Late Precambrian rocks of the Wasatch Range and Great Salt Lake area in northern Utah

Of particular importance in interpreting the origin of the conglomeratic subgraywackes is the presence or absence of a striated glacial pavement on the underlying quartzites. Such a pavement, which was carefully searched for, was

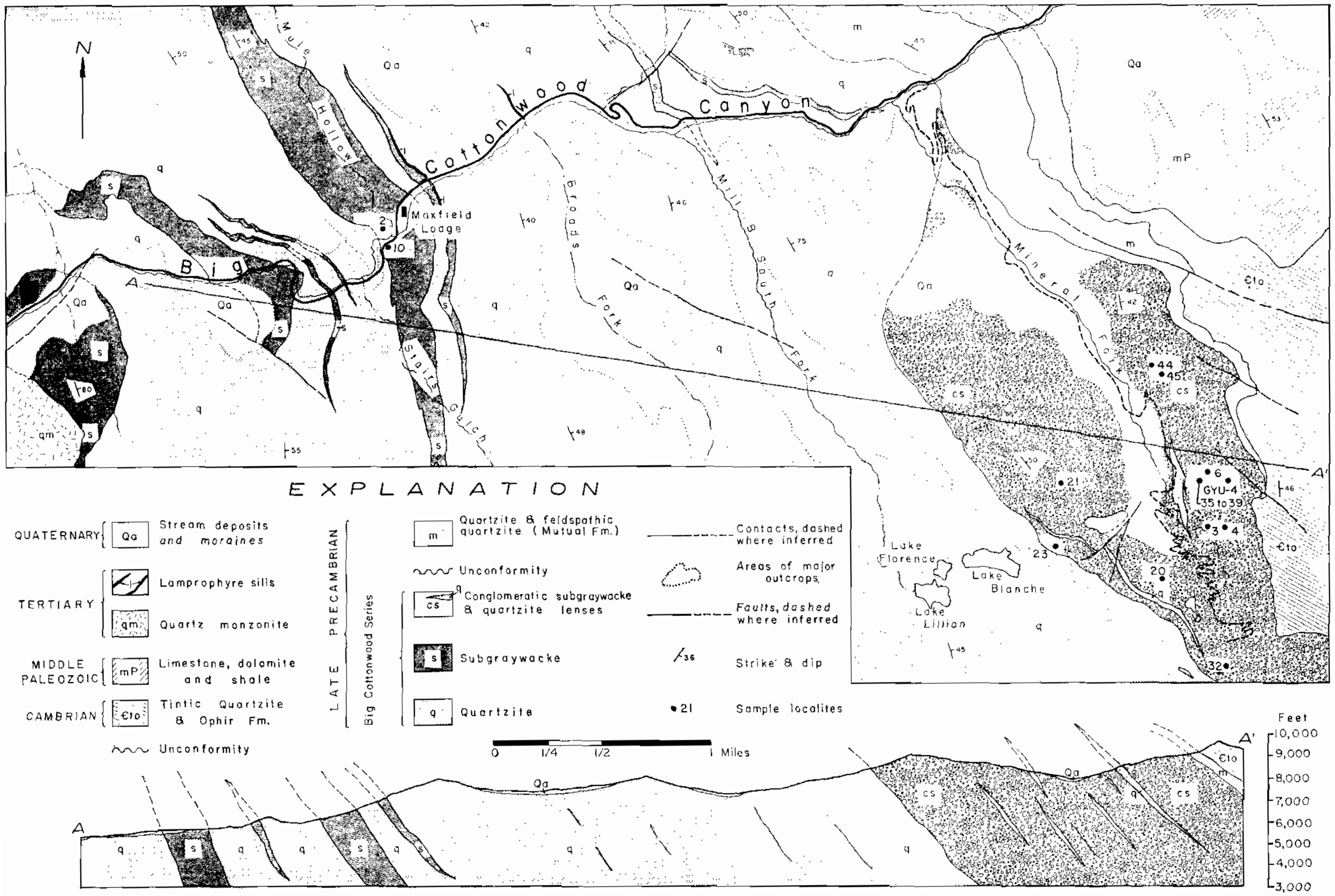



Figure 2. Geologic map and cross section of the Big Cottonwood Canyon area in the central Wasatch Range, northern Utah

not found by the author. The lower contacts of the lowest conglomeratic subgraywacke units, one of which is well exposed east of Lake Blanche, are commonly gradational with the underlying quartzites over distances of about one meter.

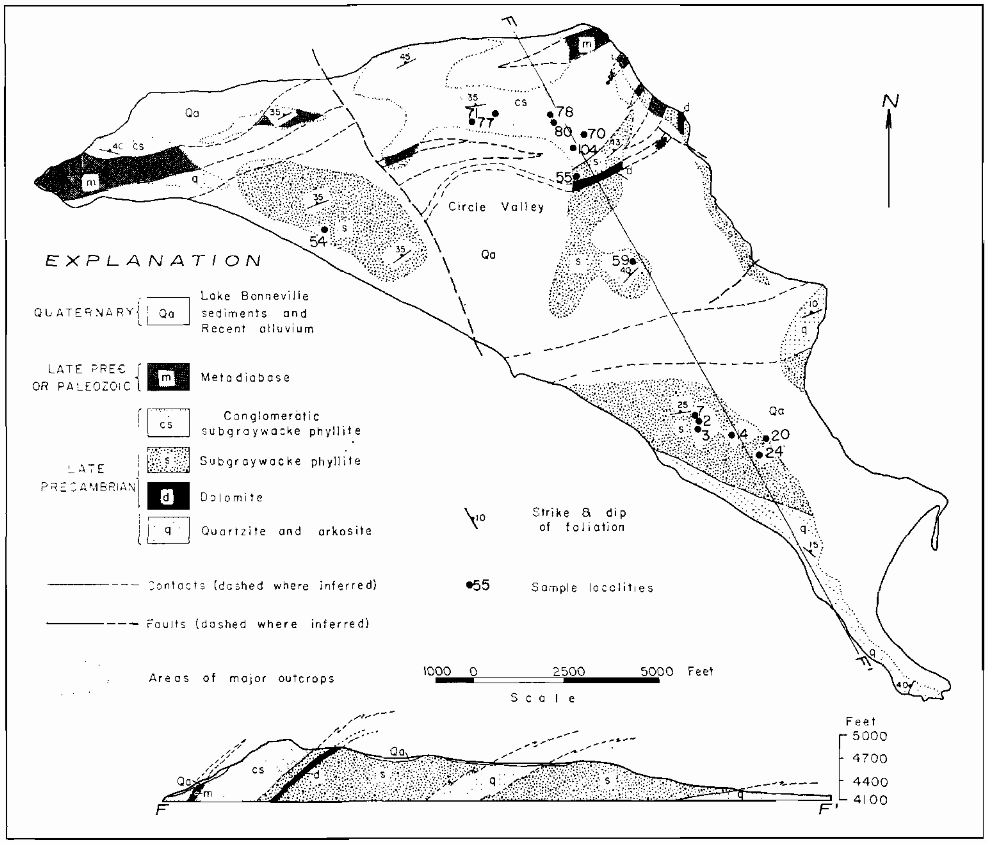



Figure 3. Geologic map and cross section of Fremont Island in Great Salt Lake, northern Utah

*Fremont Island*

Late Precambrian rocks are also well exposed on Fremont Island in Great Salt Lake (Fig. 3) where they are metamorphosed to the greenschist facies. Most of the metamorphosed subgraywackes and conglomeratic subgraywacke matrices are distinctly phyllitic or schistose and hence are mapped as subgraywacke and conglomeratic subgraywacke phyllites. These rocks, which were first described by Eardley and Hatch (1940, p. 802–804), dip north to northwest and individual beds are commonly folded on scales of few centimeters to several meters. About 2300 m of late Precambrian rocks are exposed on the island; however, because neither the top nor the base of the section is exposed, the maximum thickness cannot be determined. Metadiabase sills and dikes occur in the conglomeratic subgraywackes near the top of the section and one minor dolomite unit was found on the north-central part of the island.

The conglomeratic subgraywacke mass on Fremont Island is composed of many individual phyllitic units ranging from 0.5 to 50 m thick; unlike the conglomeratic subgraywacke mass in the Big Cottonwood Canyon area, it contains very few interbedded subgraywacke units and no quartzite units. A maximum thickness of about 900 m is exposed on the northern end of the island. As in the Big Cottonwood Canyon area, a striated pavement was not found. The lower contact of the conglomeratic subgraywacke mass is gradational with the underlying subgraywacke phyllites over distances of several meters.

Similar subgraywackes and conglomeratic subgraywackes metamorphosed to the greenschist facies occur on Antelope Island and Carrington Island in Great Salt Lake, at Little Mountain (Blackwelder, 1932), in the Sheeprock Mountains (Cohenour, 1959), northern Wasatch Mountains (Eardley and Hatch, 1940), northern Bannock Range (in southern Idaho), and Rock Canyon area east of Utah Lake.

### *The Southern Promontory Range*

A minimum of 3300 m of late Precambrian and Cambrian metamorphosed clastic sedimentary rocks occur in the southern Promontory Range. The post-Precambrian geology of the range has been described and mapped by Olson (1956); the author mapped only the southern end of the range where the late Precambrian rocks are exposed (Fig. 4). These rocks are composed dominantly of quartzite and lesser amounts of subgraywacke phyllite. Minor metadiabase dikes and sills and very minor dolomite beds were also found in the area.

The lower contact of this sequence is not exposed and the upper contact is gradational with the overlying Cambrian quartzites and quartzitic phyllites (see cross section PP′ in Fig. 4). As one goes upwards in the section, thin-bedded quartzites and subgraywackes gradually decrease in abundance and more massive, often cross-bedded quartzites increase. Such massive quartzites are common in the Lower and Middle Cambrian of the Cordilleran region. The gradual change in lithologic association in going from the late Precambrian to the lower Paleozoic in this area appears to reflect gradual changes in environments and mechanics of sedimentation and perhaps source areas. A Cambrian age is not clearly defined in the southern Promontory Range until the Middle Cambrian trilobite fauna appears in the Pioche Formation. As on Fremont Island, the section is metamorphosed to the greenschist facies and the effects of the metamorphism extend well into the overlying Paleozoic section.

## ANALYTICAL METHODS

Modal analyses were made by counting from 1000 to 2500 points on standard 20 x 30 mm thin sections. Traverses were chosen at random (normal to bedding or foliation, if present) across the thin sections. In analyzing conglomeratic subgraywackes, rock fragments >1 cm in size were not counted. It was often impossible to identify the fine, intimately intergrown minerals in many of the rocks studied, thus necessitating combined categories in the modal analyses (namely quartz or plagioclase, mica or chlorite). Staining facilitated identification of feldspar grains, and opaque minerals were identified under reflected light. Most mineral identifications were checked by X-ray diffraction.

In order to approximate the mineralogical hand specimen uniformity of the subgraywackes and conglomeratic subgraywackes, six thin sections from each of three hand specimens (approximately 3 by 5 by 10 cm in size) were analyzed counting 1000 points per section. All sections were made at approximately right angles to the bedding and the remaining rock chips were carefully cleaned and set aside for X-ray fluorescent analyses. The results are shown in Table 1. The relatively large mean deviations of quartz, plagioclase, biotite, chlorite, and muscovite modes are due to a combination of large hand specimen variability and misidentification of fine-grained minerals. The fact that the relative deviations of the quartz + plagioclase and mica + chlorite modes in subgraywacke F-3 are smaller than in the conglomeratic subgraywackes (GYU-4, LM-5) is related chiefly to nonuniform distribution of rock fragments in the conglomeratic subgraywackes which were counted in the modal analyses (and in all subsequent modal analyses reported in this investigation) according to their constituent minerals.

Seven major (Si, Al, Ti, Ca, K, Mn, and Fe) and four trace (Rb, Sr, Zr, and Ni) elements were determined in subgraywackes and conglomeratic subgraywackes by nondestructive X-ray fluorescence analysis. All samples were ground to approximately the same size (~20 microns) and pelletized in a procedure similar to that described by Baird (1961). A calibration curve was constructed for each major element using the mean values of 8 wet chemical analyses of each two samples (F-3 and GYU-4). The four trace elements were determined in the manner proposed by Reynolds (1963) by estimating mass absorption coefficients of each sample with Compton scattered $MoK_\alpha$ radiation and using G-1 (for Rb, Sr, and Zr) or W-1 (for Ni) as standards. The precision of pellet reproducibility was less than 3 percent for all elements. The estimated over-all accuracy

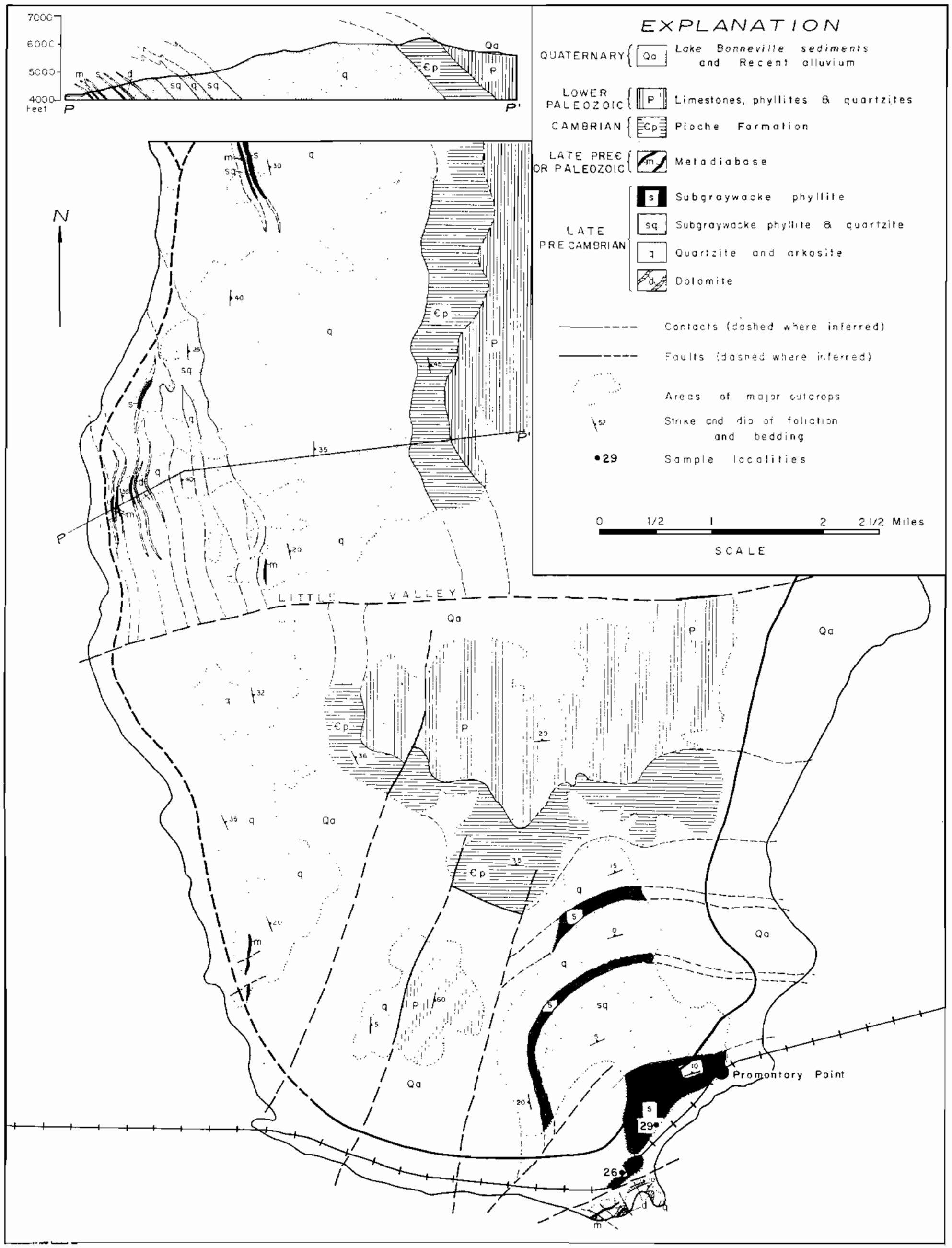



Figure 4. Geologic map and cross section of the southern Promontory Range, northern Utah. Post-Precambrian geology after Olson, 1956

TABLE 1. PRECISION OF MODAL ANALYSES

| | Thin section* 1 | 2 | 3 | 4 | 5 | 6 | $\bar{x}$† | S | C |
|---|---|---|---|---|---|---|---|---|---|
| **Subgraywacke phyllite F–3** | | | | | | | | | |
| Quartz | 46.5 | 45.4 | 45.4 | 46.3 | 44.9 | 47.1 | 45.9 | 0.8 | 1.7 |
| Plagioclase | 2.8 | 1.9 | 2.1 | 2.7 | 2.5 | 1.8 | 2.3 | 0.4 | 17.4 |
| Quartz or plagioclase | 5.2 | 7.2 | 7.3 | 4.8 | 8.7 | 4.8 | 6.3 | 1.6 | 25.4 |
| Total quartz + plagioclase | 54.5 | 54.5 | 54.8 | 52.8 | 56.1 | 53.7 | 54.5 | 0.9 | 1.7 |
| Muscovite‡ | 22.6 | 20.5 | 18.5 | 21.8 | 21.4 | 23.2 | 21.3 | 1.7 | 8.0 |
| Chlorite | 13.1 | 10.5 | 6.7 | 12.9 | 17.1 | 12.9 | 12.2 | 3.4 | 27.8 |
| Mica or chlorite | 6.5 | 11.1 | 16.7 | 8.5 | 1.6 | 7.2 | 8.6 | 5.0 | 58.1 |
| Total mica + chlorite | 42.2 | 42.1 | 41.9 | 43.2 | 40.1 | 43.3 | 42.1 | 1.2 | 2.9 |
| Chloritoid | 0.2 | 0.1 | 0.2 | 0.2 | 0.3 | 0.2 | 0.2 | <0.1 | <50.0 |
| Carbonaceous matter + magnetite + ilmenite | 2.6 | 3.0 | 2.7 | 3.5 | 3.2 | 2.4 | 2.9 | 0.4 | 13.8 |
| Limonite + hematite | 0.2 | 0.1 | 0.1 | 0.2 | 0.1 | 0.3 | 0.2 | <0.1 | <50.0 |
| Apatite | tr | tr | tr | tr | 0.1 | 0.1 | tr | .. | .. |
| Zircon | tr | tr | .. | tr | .. | tr | tr | .. | .. |
| **Conglomeratic subgraywacke GYU–4** | | | | | | | | | |
| Quartz | 20.5 | 15.3 | 22.1 | 19.5 | 13.9 | 18.0 | 18.2 | 3.1 | 17.0 |
| Plagioclase | 11.0 | 5.7 | 8.2 | 9.1 | 5.9 | 7.6 | 7.9 | 2.0 | 25.3 |
| Quartz or plagioclase | 6.9 | 7.9 | 10.3 | 1.5 | 24.6 | 17.8 | 11.5 | 8.3 | 72.1 |
| Total quartz + plagioclase | 38.4 | 39.3 | 40.6 | 30.1 | 44.4 | 43.4 | 39.3 | 5.0 | 12.7 |
| Biotite | 15.3 | 14.3 | 9.7 | 10.5 | 11.1 | 17.3 | 13.0 | 3.0 | 23.1 |
| Muscovite | 3.2 | 2.1 | 3.5 | 1.8 | 1.7 | 3.3 | 2.6 | 0.8 | 30.8 |
| Chlorite | 1.5 | 0.8 | 1.7 | 0.9 | 1.0 | 1.4 | 1.2 | 0.4 | 33.3 |
| Mica or chlorite | 34.1 | 36.7 | 37.4 | 50.0 | 34.3 | 27.8 | 36.8 | 8.2 | 22.2 |
| Total mica + chlorite | 54.1 | 53.9 | 52.2 | 63.2 | 48.1 | 49.8 | 53.6 | 7.9 | 14.7 |
| Microcline | 0.2 | 0.5 | 1.1 | 0.9 | 0.3 | 0.7 | 0.6 | 0.4 | 66.7 |
| Calcite | 1.4 | 1.2 | 0.8 | 0.1 | 1.3 | 1.0 | 1.0 | 0.4 | 40.0 |
| Carbonaceous matter + magnetite + ilmenite | 5.7 | 5.0 | 5.0 | 5.5 | 5.7 | 5.0 | 5.3 | 0.4 | 7.6 |
| Limonite + hematite | tr | .. | tr | 0.1 | tr | .. | .. | .. | .. |
| Apatite | .. | .. | tr | .. | .. | tr | .. | .. | .. |
| Zircon | tr | tr | tr | tr | tr | tr | tr | .. | .. |
| **Conglomeratic subgraywacke phyllite LM–5** | | | | | | | | | |
| Quartz | 41.0 | 30.2 | 22.8 | 28.8 | 28.6 | 37.5 | 31.5 | 6.6 | 21.0 |
| Plagioclase | 5.2 | 8.0 | 16.9 | 15.3 | 14.1 | 9.2 | 11.5 | 4.6 | 40.0 |
| Quartz or plagioclase | 12.6 | 21.9 | 27.0 | 20.0 | 18.8 | 16.9 | 19.5 | 4.8 | 24.6 |
| Total quartz + plagioclase | 58.8 | 60.1 | 66.7 | 64.1 | 61.5 | 63.6 | 62.5 | 2.9 | 4.6 |
| Biotite | 11.5 | 9.3 | 19.1 | 16.2 | 10.8 | 16.2 | 13.9 | 3.9 | 28.0 |
| Muscovite | 6.3 | 5.7 | 1.2 | 4.5 | 5.2 | 3.1 | 4.3 | 1.9 | 44.1 |
| Chlorite | 1.5 | 2.1 | 1.1 | 2.9 | 1.0 | 0.9 | 1.6 | 0.8 | 50.0 |
| Mica or chlorite | 19.6 | 21.2 | 9.6 | 10.2 | 20.1 | 14.0 | 15.8 | 5.2 | 32.9 |
| Total mica + chlorite | 38.9 | 38.3 | 31.0 | 33.8 | 37.1 | 34.2 | 35.6 | 3.0 | 8.4 |
| Calcite | 0.5 | 0.2 | 0.6 | 0.5 | 0.1 | 0.6 | 0.4 | 0.2 | 50.0 |
| Carbonaceous matter + magnetite + ilmenite | 1.5 | 1.3 | 1.6 | 1.4 | 1.1 | 1.3 | 1.3 | 0.2 | 15.4 |
| Limonite + hematite | tr | tr | tr | tr | tr | tr | tr | .. | .. |
| Apatite | tr | tr | tr | tr | tr | tr | tr | .. | .. |

* 1,000 points per thin section

† $\bar{x}$–mean; S–standard deviation; C–relative deviation (percent)

‡ Including paragonite

which for the major elements is principally controlled by the accuracy of the wet chemical analyses of the two samples from which the calibration curves were constructed is given in Table 2.

Modal and major and trace element analyses of typical subgraywackes and conglomeratic subgraywackes from both unmetamorphosed and metamorphosed late Precambrian sections in the northeastern Great Basin are given in Tables 3 and 4. The localities of samples collected from the Big Cottonwood Canyon area,

Table 2. Estimated Accuracy of X-ray Fluorescence Data (percent)

| | |
|---|---|
| $SiO_2$ | 3 |
| $Al_2O_3$ | 8 |
| $TiO_2$ | 5 |
| CaO | 5 |
| $K_2O$ | 8 |
| Total Fe as $Fe_2O_3$ | 7 |
| MnO | 10 |
| Rb | 5 |
| Sr | 5 |
| Zr | 3 |
| Ni | 5 |

Fremont Island, and the southern Promontory Range are indicated in Figures 2, 3, and 4.

## QUARTZITE

Quartzite is by far the most abundant rock type in the late Precambrian of northern Utah, constituting from 70 to 80 percent of most exposed sections. It occurs as thick and commonly cross-bedded units composed of continuous quartzite sections of $\geq$5,000 m and as thin-bedded laminae intimately interbedded and often gradational with subgraywacke. This latter type of quartzite will be considered with the subgraywackes.

Quartzites of the first type, in which individual beds range in thickness from about 10 cm to 4 m (averaging about 20 cm), are almost identical to their counterparts in the lower Paleozoic rocks of the Great Basin and adjacent areas in the Rocky Mountains. This, in fact, is a major factor contributing to the difficulty in defining the Precambrian-Cambrian boundary in most areas where an obvious unconformity is nonexistent (for example in the southern Promontory Range).

Thicker, cross-bedded quartzites also occur interbedded in the subgraywacke and conglomeratic subgraywacke sections. A few such larger units are shown on the geologic map of the Big Cottonwood Canyon area (Fig. 2) and in the sketches of the Mineral Fork conglomeratic subgraywackes (Fig. 5). With exception of these quartzite units, which have obvious elongate lenticular shapes, most quartzite units have very similar thicknesses for great distances. Several units were followed laterally for distances greater than 200 m in the Lake Blanche area of Big Cottonwood Canyon without perceptible changes in thickness.

The thicker quartzite beds, unlike the thin-bedded varieties associated with the subgraywackes, do not have fine-grained matrices but contain quartz grains of very limited size ranges (usually from 0.8 to 1.5 mm). The original grain shapes, as preserved by quartz overgrowths, are usually subangular to subrounded and in this way are different from those found in the subgraywackes and conglomeratic subgraywackes.

Nearly all of the quartzite units observed by the author were cross-bedded. Although festoon cross-bedding occurs in some of the thinner units, planar cross-bedding is most abundant. The individual cross-bedded laminae range in length from 0.1 to 1 m and are usually from 0.5 to 5 cm thick; they dip from 25 to 35 degrees in directions indicating westerly to southwesterly transport. Mud cracks, generally filled with fine-grained clastic material, are common on the upper surfaces of some quartzite units. Questionable ripple marks and worm burrows are also occasionally found in these rocks.

Plagioclase and microcline constitute a small amount of many of the quartzites. Plagioclase ranges from $An_{25}$ to $An_{45}$ in the unmetamorphosed samples and from $An_{28}$ to $An_{33}$ in samples metamorphosed to the greenschist facies. Mica (chiefly muscovite) and chlorite also form minor yet ubiquitous components in most of the quartzites. As in the subgraywackes, they commonly penetrate and embay clastic quartz grains suggesting a diagenetic or metamorphic origin. Pyrite, probably of metamorphic origin, occurs in some of the metamorphosed quartzites.

## SUBGRAYWACKE AND CONGLOMERATIC SUBGRAYWACKE

### *Primary Textures and Structures*

The late Precambrian subgraywackes of this region are massive, well indurated, commonly graded, greenish-black to charcoal-black quartzitic sandstones, siltstones, and associated

TABLE 3. MODAL AND X-RAY FLUORESCENT CHEMICAL ANALYSES OF THE SUBGRAYWACKES

| | Unmetamorphosed samples | | | | | | | | Metamorphosed samples | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | BC–2 | BC–3 | BC–6 | BC–10b | BC–20b | BC–44 | $\bar{x}$[††] | M.D. | F–3 | F–7 | F–14 | F–24 | F–54 | F–59 | PP–26 | PP–29 | $\bar{x}$ | M.D. |
| Modal analyses (volume percent) | | | | | | | | | | | | | | | | | | |
| Quartz | 10.5 | 6.8 | 31.5 | 14.1 | 8.0 | 11.3 | 13.7 | 6.1 | 45.9 | 31.2 | 8.2 | 25.5 | 21.2 | 15.3 | 10.1 | 14.8 | 21.5 | 9.5 |
| Plagioclase | 6.1 | 1.2 | 2.6 | 1.5 | tr | 8.2 | 3.3 | 2.6 | 2.3 | 5.1 | 0.9 | 8.6 | 6.5 | 2.1 | 0.2 | 4.8 | 3.8 | 2.4 |
| Quartz or plagioclase | 7.2 | 4.5 | 23.9 | 4.0 | 7.1 | 3.0 | 8.3 | 5.2 | 6.3 | 24.1 | 5.5 | 5.2 | 9.8 | 19.8 | 8.0 | 6.9 | 10.7 | 5.6 |
| Total quartz + plagioclase | 23.8 | 12.5 | 58.0 | 19.6 | 15.1 | 22.5 | 25.3 | 11.0 | 54.4 | 60.4 | 14.6 | 39.3 | 37.5 | 37.2 | 18.3 | 26.5 | 36.0 | 12.2 |
| Biotite | tr | .. | .. | .. | tr | .. | .. | .. | .. | 1.2 | .. | .. | .. | .. | .. | .. | .. | .. |
| Muscovite* | 3.1 | tr | tr | 1.5 | 1.2 | 3.1 | 1.5 | 1.1 | 21.3 | 20.5 | 2.3 | 17.9 | 13.2 | 5.7 | 8.1 | 18.5 | 13.4 | 6.1 |
| Chlorite | .. | .. | .. | tr | 5.1 | 5.8 | 1.8 | 2.4 | 12.2 | tr | 5.1 | 3.1 | 8.1 | 14.2 | 0.5 | 5.1 | 6.0 | 4.1 |
| Mica or chlorite | 66.4 | 8.2 | 13.3 | 64.7 | 76.9 | 45.1 | 45.8 | 23.6 | 8.6 | 9.3 | 70.7 | 35.9 | 31.7 | 31.2 | 67.6 | 43.0 | 37.3 | 17.4 |
| Total mica + chlorite | 69.5 | 8.2 | 13.3 | 66.2 | 83.2 | 54.0 | 49.1 | 25.5 | 42.1 | 31.0 | 78.1 | 56.9 | 53.0 | 51.1 | 76.2 | 66.6 | 56.7 | 12.6 |
| Microcline | tr | tr | 4.2 | tr | tr | 1.9 | 1.0 | 1.3 | .. | .. | .. | .. | .. | .. | tr | .. | .. | .. |
| Chloritoid | 2.2 | .. | .. | 2.7 | .. | .. | 0.8 | 1.1 | 0.2 | 6.0 | 1.8 | 3.5 | 5.8 | 9.3 | 0.1 | 0.1 | 3.4 | 2.8 |
| Siderite | tr | 75.5 | 21.0 | 5.1 | 0.3 | 13.5 | 19.2 | 19.3 | .. | .. | .. | .. | .. | .. | .. | .. | .. | .. |
| Carbonaceous matter + magnetite + ilmenite | 4.3 | 3.9 | 3.3 | 6.3 | 1.2 | 8.0 | 4.5 | 1.8 | 2.9 | 2.5 | 4.9 | 0.2 | 3.1 | 2.3 | 5.3 | 5.3 | 3.3 | 1.4 |

| | | | | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Limonite + hematite | tr | .. | .. | tr | .. | .. | .. | .. | 0.2 | tr | 0.5 | .. | 0.1 | tr | .. | .. | 0.1 | 0.1 |
| Pyrite | .. | .. | .. | .. | .. | .. | .. | .. | .. | tr | .. | tr | 0.3 | tr | tr | 1.3 | 0.2 | 0.3 |
| Apatite | tr | .. | tr | tr | tr | tr | .. | .. | tr | tr | tr | .. | .. | .. | tr | tr | .. | .. |
| Zircon | .. | .. | .. | tr | tr | .. | .. | .. | 0.1 | 0.1 | .. | tr | 0.1 | tr | tr | tr | .. | .. |
| Other† | tr | tr | tr | tr | tr | tr | .. | .. | .. | tr | tr | .. | tr | tr | .. | .. | .. | .. |
| Major elements as oxides (weight percent)‡ | | | | | | | | | | | | | | | | | | |
| $SiO_2$ | 61.24 | 14.82 | 69.09 | 65.81 | 69.72 | 58.32 | 56.50 | 13.89 | 64.77 | 79.91 | 58.32 | 77.49 | 75.79 | 74.33 | 60.40 | 76.63 | 70.96 | 9.79 |
| $TiO_2$ | 0.94 | 0.58 | 0.49 | 1.12 | 0.87 | 0.67 | 0.78 | 0.23 | 0.84 | 0.49 | 1.10 | 0.61 | 1.06 | 1.28 | 0.96 | 0.59 | 0.87 | 0.23 |
| $Al_2O_3$ | 16.78 | 1.53 | 9.99 | 16.62 | 14.59 | 10.83 | 11.72 | 4.27 | 20.69 | 18.62 | 23.09 | 16.46 | 15.21 | 19.39 | 15.73 | 17.15 | 18.29 | 2.16 |
| CaO | 1.28 | 1.03 | 2.07 | 0.47 | 1.13 | 1.32 | 1.22 | 0.34 | 0.22 | 0.14 | 0.72 | 0.16 | 0.53 | 0.18 | 0.17 | 0.20 | 0.29 | 0.17 |
| $K_2O$ | 2.43 | 0.82 | 2.33 | 3.13 | 2.26 | 2.67 | 2.27 | 0.49 | 2.63 | 2.31 | 3.02 | 2.59 | 2.54 | 3.04 | 3.11 | 2.97 | 2.78 | 0.26 |
| MnO | 0.07 | 0.09 | 0.15 | 0.06 | 0.08 | 0.19 | 0.11 | 0.04 | 0.02 | 0.01 | 0.05 | 0.10 | 0.07 | 0.03 | 0.03 | 0.08 | 0.05 | 0.03 |
| $Fe_2O_3$** | 9.21 | 47.71 | 5.19 | 5.35 | 7.58 | 12.98 | 14.67 | 11.01 | 6.36 | 3.01 | 5.15 | 5.21 | 4.60 | 4.75 | 6.60 | 6.15 | 5.23 | 0.86 |
| Trace elements (ppm)‡ | | | | | | | | | | | | | | | | | | |
| Rb | 118 | 129 | 82 | 137 | 122 | 115 | 117 | 13 | 116 | 112 | 182 | 133 | 127 | 178 | 170 | 143 | 145 | 24 |
| Sr | 93 | 81 | 78 | 111 | 69 | 84 | 86 | 11 | 112 | 48 | 247 | 39 | 34 | 118 | 51 | 74 | 90 | 51 |
| Zr | 250 | 138 | 250 | 256 | 268 | 196 | 226 | 40 | 488 | 475 | 217 | 449 | 494 | 290 | 156 | 123 | 337 | 140 |
| Ni | 54 | 72 | 21 | 55 | 59 | 28 | 48 | 16 | 33 | 28 | 29 | 30 | 40 | 38 | 35 | 42 | 34 | 4 |

* Including paragonite in the metamorphosed samples
† Chiefly calcite, sphene and rutile
‡ Each value represents the mean of three or more pellets
** Total Fe as $Fe_2O_3$
†† $\bar{x}$–mean; M.D.–mean deviation

Sample localities:
BC–Big Cottonwood Canyon area (Fig. 2)
F–Fremont Island (Fig. 3)
PP–Promontory Point (Fig. 4)

TABLE 4. MODAL AND X-RAY FLUORESCENT CHEMICAL ANALYSES OF THE CONGLOMERATIC SUBGRAYWACKES

| | Unmetamorphosed samples | | | | | | | Metamorphosed samples | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | BC-4 | GYU-4 | BC-21 | BC-38 | BC-39 | x̄** | M.D. | LM-5 | RC-4 | P-8 | Sh-3 | F-70 | F-71 | LM-23 | AI-2 | x̄ | M.D. |
| | | | | | | | | Modal analyses (volume percent) | | | | | | | | | |
| Quartz | 30.6 | 18.2 | 36.8 | 49.9 | 35.3 | 34.1 | 7.8 | 31.5 | 46.6 | 20.1 | 23.9 | 36.5 | 26.9 | 38.7 | 24.1 | 31.0 | 7.3 |
| Plagioclase | 6.7 | 7.9 | 9.1 | 2.0 | 9.1 | 7.0 | 2.1 | 11.5 | 1.1 | 7.0 | 0.1 | 3.8 | 7.6 | 4.1 | 2.8 | 4.8 | 3.0 |
| Quartz or plagioclase | 3.9 | 11.5 | 5.5 | 10.5 | 5.0 | 7.3 | 3.0 | 19.5 | 7.1 | 7.0 | .. | 7.8 | 9.4 | 11.6 | 2.0 | 8.1 | 3.1 |
| Total quartz + plagioclase | 41.2 | 39.3 | 51.4 | 62.4 | 49.4 | 48.7 | 6.8 | 62.5 | 54.8 | 34.1 | 24.0 | 48.1 | 43.9 | 54.4 | 28.9 | 43.8 | 11.1 |
| Biotite | 5.1 | 13.0 | 9.3 | .. | .. | 5.5 | 4.5 | 13.9 | tr | tr | 0.3 | tr | tr | 2.3 | 0.3 | 2.1 | 2.0 |
| Muscovite | 3.5 | 2.6 | 4.8 | 4.2 | 3.2 | 3.7 | 0.7 | 4.3 | 12.1 | 3.5 | 6.3 | 9.7 | 5.3 | 9.6 | 49.5 | 12.5 | 9.2 |
| Chlorite | 7.2 | 1.2 | 3.2 | 12.1 | 10.1 | 6.8 | 3.6 | 1.6 | 5.3 | 18.3 | 35.8 | 0.8 | 1.9 | 8.9 | .. | 9.1 | 7.9 |
| Mica or chlorite | 31.2 | 36.8 | 25.6 | 9.9 | 18.7 | 24.4 | 8.1 | 15.8 | 20.9 | 28.3 | 27.0 | 15.9 | 41.3 | 22.5 | 4.5 | 22.0 | 9.0 |
| Total mica + chlorite | 47.0 | 53.6 | 42.9 | 26.2 | 32.0 | 40.3 | 9.0 | 35.6 | 38.3 | 50.1 | 69.4 | 26.4 | 48.5 | 43.4 | 54.3 | 45.8 | 9.8 |
| Microcline | 4.1 | 0.6 | 2.5 | 1.2 | 7.4 | 3.2 | 2.1 | .. | 0.2 | .. | .. | .. | .. | .. | 4.3 | 0.6 | 0.5 |
| Calcite | .. | 1.0 | .. | .. | tr | .. | .. | 0.4 | tr | 10.8 | 4.4 | 15.8 | .. | tr | .. | 5.0 | 2.7 |
| Dolomite | 2.8 | .. | tr | 8.7 | 6.1 | 3.5 | 3.1 | .. | 5.8 | .. | .. | .. | .. | .. | .. | .. | .. |
| Siderite | tr | .. | 0.2 | tr | tr | .. | .. | .. | .. | .. | .. | .. | .. | .. | .. | .. | .. |
| Carbonaceous matter + magnetite + ilmenite | 4.7 | 5.3 | 2.9 | 1.4 | 4.8 | 3.8 | 1.3 | 1.4 | 0.9 | 3.2 | 1.0 | 1.3 | 5.5 | 2.2 | 12.0 | 3.4 | 2.6 |

| | | | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Limonite + hematite | .. | tr | tr | .. | .. | .. | .. | tr | tr | tr | tr | 8.2 | 1.9 | 0.1 | 0.2 | 1.3 | 1.2 |
| Apatite | .. | tr | .. | tr | .. | .. | .. | tr | .. | .. | .. | tr | .. | .. | tr | tr | .. |
| Zircon | tr | tr | .. | .. | .. | .. | .. | .. | tr | tr | tr | 0.1 | tr | 0.2 | tr | tr | .. |
| Other* | .. | .. | .. | .. | tr | .. | .. | tr | .. | 1.7 | 1.2 | .. | tr | .. | tr | tr | .. |
| Major elements as oxides (weight percent)† | | | | | | | | | | | | | | | | | |
| $SiO_2$ | 69.76 | 66.85 | 78.69 | 73.71 | 80.29 | 73.86 | 4.50 | 69.53 | 75.41 | 62.14 | 67.42 | 88.53 | 66.44 | 85.48 | 68.93 | 72.99 | 7.62 |
| $TiO_2$ | 0.54 | 0.57 | 0.66 | 1.01 | 0.49 | 0.65 | 0.14 | 0.82 | 0.44 | 1.70 | 0.82 | 0.57 | 0.67 | 0.65 | 0.53 | 0.78 | 0.26 |
| $Al_2O_3$ | 10.62 | 12.24 | 10.94 | 11.25 | 10.05 | 11.02 | 0.58 | 13.35 | 8.90 | 10.78 | 13.28 | 14.80 | 14.53 | 14.17 | 13.65 | 12.93 | 1.55 |
| CaO | 1.31 | 1.69 | 1.06 | 0.73 | 1.01 | 1.16 | 0.27 | 0.39 | 2.48 | 4.39 | 2.90 | 1.62 | 0.49 | 0.35 | 0.21 | 1.60 | 1.24 |
| $K_2O$ | 2.97 | 2.64 | 2.84 | 2.73 | 2.30 | 2.70 | 0.18 | 3.49 | 2.62 | 2.93 | 3.07 | 3.44 | 4.40 | 3.21 | 5.38 | 3.57 | 0.46 |
| MnO | 0.06 | 0.08 | 0.06 | 0.05 | 0.20 | 0.09 | 0.04 | 0.10 | 0.04 | 0.10 | 0.16 | 0.18 | 0.06 | 0.07 | 0.03 | 0.09 | 0.04 |
| $Fe_2O_3$‡ | 5.87 | 9.71 | 5.54 | 5.16 | 5.11 | 6.28 | 1.37 | 5.56 | 3.28 | 6.66 | 5.81 | 3.70 | 5.79 | 4.51 | 6.02 | 5.17 | 1.00 |
| Trace elements (ppm)† | | | | | | | | | | | | | | | | | |
| Rb | 106 | 100 | 89 | 92 | 85 | 94 | 7 | 141 | 74 | 129 | 232 | 139 | 186 | 152 | 198 | 156 | 37 |
| Sr | 81 | 79 | 42 | 27 | 47 | 55 | 20 | 70 | 15 | 352 | 106 | 112 | 52 | 94 | 21 | 103 | 66 |
| Zr | 187 | 213 | 194 | 255 | 267 | 223 | 30 | 393 | 175 | 531 | 442 | 474 | 3888 | 666 | 299 | 421 | 107 |
| Ni | 22 | 30 | 31 | 40 | 16 | 28 | 7 | 31 | 31 | 57 | 36 | 45 | 38 | 18 | 47 | 38 | 6 |

* Primarily rutile and sphene; actinolite and epidote in samples Sh–3 and P–8
† Each value represents the mean of three or more pellets
‡ Total Fe as $Fe_2O_3$
** $\bar{x}$–mean; M.D.–mean deviation

Sample localities:
BC, GYU–Big Cottonwood Canyon area (Fig. 2)
LM–Little Mountain
RC–Rock Canyon, Central Wasatch Mountains
AI–Antelope Island
F–Fremont Island (Fig. 3)
P–Northern Bannock Range
Sh–Sheeprock Mountains

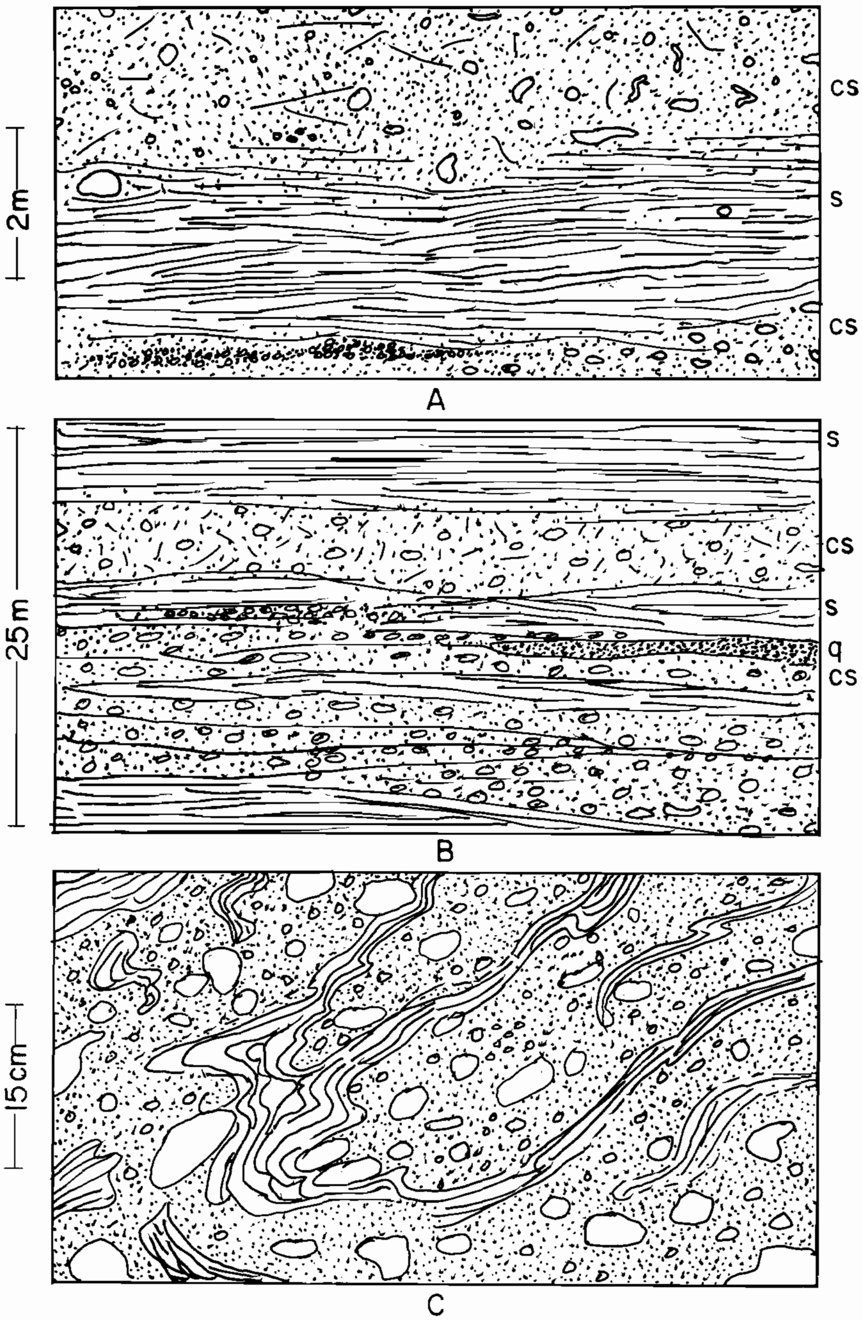
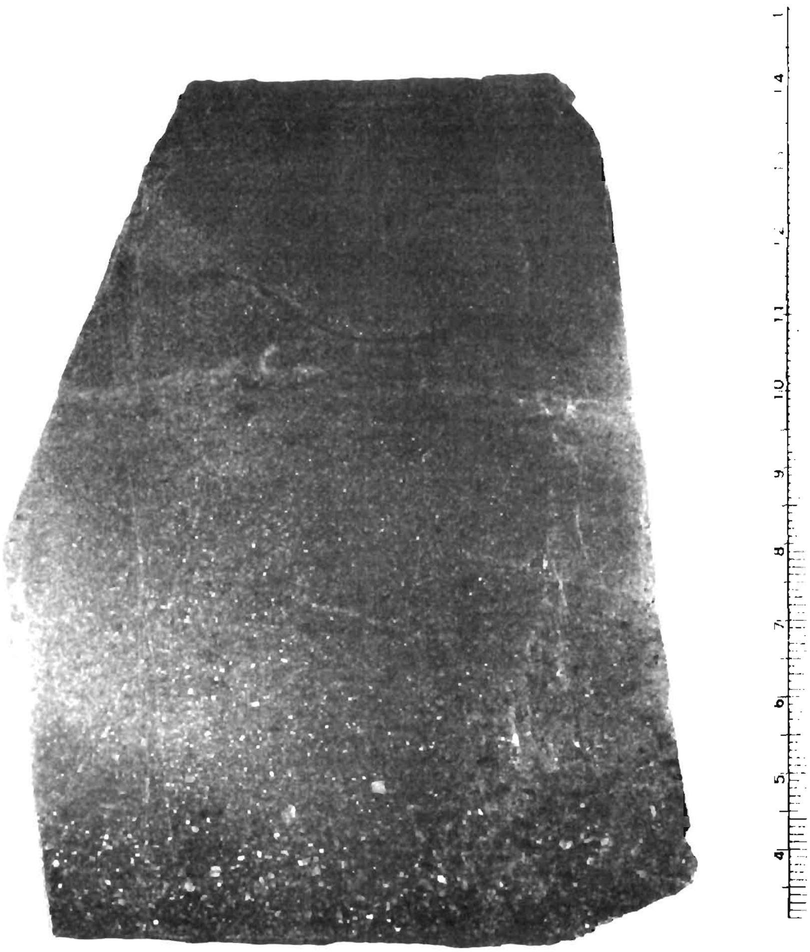



Figure 5. Geologic field sketches of (A) interbedded and gradational conglomeratic subgraywacke (cs) and subgraywacke (s) in Mineral Fork, Big Cottonwood Canyon area; (B) sketch of a 25 m section in Mineral Fork, showing relationships between subgraywacke (s), conglomeratic subgraywacke (cs) and massive quartzite (q); and (C) twisted and bent subgraywacke fragments in a typical conglomeratic subgraywacke unit in Mineral Fork

| | | | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Limonite + hematite | .. | tr | tr | .. | .. | .. | .. | tr | tr | tr | tr | 8.2 | 1.9 | 0.1 | 0.2 | 1.3 | 1.2 |
| Apatite | .. | tr | .. | tr | .. | .. | .. | tr | .. | .. | .. | tr | .. | .. | tr | tr | .. |
| Zircon | tr | tr | .. | .. | .. | .. | .. | .. | tr | tr | tr | 0.1 | tr | 0.2 | tr | tr | .. |
| Other* | .. | .. | .. | .. | tr | .. | .. | tr | .. | 1.7 | 1.2 | .. | tr | .. | tr | tr | .. |
| Major elements as oxides (weight percent)† | | | | | | | | | | | | | | | | | |
| $SiO_2$ | 69.76 | 66.85 | 78.69 | 73.71 | 80.29 | 73.86 | 4.50 | 69.53 | 75.41 | 62.14 | 67.42 | 88.53 | 66.44 | 85.48 | 68.93 | 72.99 | 7.62 |
| $TiO_2$ | 0.54 | 0.57 | 0.66 | 1.01 | 0.49 | 0.65 | 0.14 | 0.82 | 0.44 | 1.70 | 0.82 | 0.57 | 0.67 | 0.65 | 0.53 | 0.78 | 0.26 |
| $Al_2O_3$ | 10.62 | 12.24 | 10.94 | 11.25 | 10.05 | 11.02 | 0.58 | 13.35 | 8.90 | 10.78 | 13.28 | 14.80 | 14.53 | 14.17 | 13.65 | 12.93 | 1.55 |
| CaO | 1.31 | 1.69 | 1.06 | 0.73 | 1.01 | 1.16 | 0.27 | 0.39 | 2.48 | 4.39 | 2.90 | 1.62 | 0.49 | 0.35 | 0.21 | 1.60 | 1.24 |
| $K_2O$ | 2.97 | 2.64 | 2.84 | 2.73 | 2.30 | 2.70 | 0.18 | 3.49 | 2.62 | 2.93 | 3.07 | 3.44 | 4.40 | 3.21 | 5.38 | 3.57 | 0.46 |
| MnO | 0.06 | 0.08 | 0.06 | 0.05 | 0.20 | 0.09 | 0.04 | 0.10 | 0.04 | 0.10 | 0.16 | 0.18 | 0.06 | 0.07 | 0.03 | 0.09 | 0.04 |
| $Fe_2O_3$‡ | 5.87 | 9.71 | 5.54 | 5.16 | 5.11 | 6.28 | 1.37 | 5.56 | 3.28 | 6.66 | 5.81 | 3.70 | 5.79 | 4.51 | 6.02 | 5.17 | 1.00 |
| Trace elements (ppm)† | | | | | | | | | | | | | | | | | |
| Rb | 106 | 100 | 89 | 92 | 85 | 94 | 7 | 141 | 74 | 129 | 232 | 139 | 186 | 152 | 198 | 156 | 37 |
| Sr | 81 | 79 | 42 | 27 | 47 | 55 | 20 | 70 | 15 | 352 | 106 | 112 | 52 | 94 | 21 | 103 | 66 |
| Zr | 187 | 213 | 194 | 255 | 267 | 223 | 30 | 393 | 175 | 531 | 442 | 474 | 3888 | 666 | 299 | 421 | 107 |
| Ni | 22 | 30 | 31 | 40 | 16 | 28 | 7 | 31 | 31 | 57 | 36 | 45 | 38 | 18 | 47 | 38 | 6 |

* Primarily rutile and sphene; actinolite and epidote in samples Sh-3 and P-8
† Each value represents the mean of three or more pellets
‡ Total Fe as $Fe_2O_3$
** $\bar{x}$-mean; M.D.-mean deviation

Sample localities:
BC, GYU-Big Cottonwood Canyon area (Fig. 2)
LM-Little Mountain
RC-Rock Canyon, Central Wasatch Mountains
AI-Antelope Island
F-Fremont Island (Fig. 3)
P-Northern Bannock Range
Sh-Sheeprock Mountains



A TYPICAL GRADED SUBGRAYWACKE BED (BC-45) FROM MINERAL FORK IN THE BIG COTTONWOOD CANYON AREA

Note the subgraywacke fragments which appear as black patches and twisted laminae. Scale in centimeters.

phyllitic shales. They grade both vertically and horizontally into thin-bedded, light-colored quartzites (containing >90 percent quartz). Commonly the lower portions of graded units should more appropriately be called quartzite (or subgraywacke quartzite) whereas their upper portions should be called subgraywacke; in other cases the distinction between quartzite and subgraywacke is quite pronounced. Sections of intimately interbedded quartzite and subgraywacke laminae (5 mm to 5 cm thick) occur in many areas. At some localities the quartzite laminae have sharp contacts with the subgraywacke laminae thus imparting a varve-like, striped appearance to the rocks. Because subgraywacke and thin-bedded quartzite are gradational and can only arbitrarily be distinguished, the term subgraywacke will be used in this investigation (unless otherwise specified) to include both subgraywacke proper and intimately associated thin-bedded quartzite. The gradational nature of these rocks strongly indicates that they were deposited by the same transporting agent.

Many individual subgraywacke units were traced horizontally, and regardless of their thickness, were found to have elongate lenticular shapes. Such shapes are dominant in graded as well as nongraded units. Subgraywackes in some areas are complexly interfingered and gradational with conglomeratic subgraywackes. Figure 5 (A and B) shows sketches of some of the field relations of these rock types that are so well exposed in Mineral Fork in the Big Cottonwood Canyon area.

The subgraywackes contain silt- to sand-size quartz and minor feldspar grains in a finer matrix of chlorite, mica, quartz, feldspar, carbonates, carbonaceous matter, and iron oxides. Their sand grains are subrounded to commonly well-rounded. Because there is a complete gradation from sand- to clay-size particles, a matrix as such, can only be arbitrarily defined. If all particles ≤0.1 mm are defined as matrix, the matrix varies in abundance from 10 to 70 percent. The chlorite, mica, and most other matrix constituents in the size fraction ≤0.1 mm (and in some larger grains) commonly penetrate and embay rounded sand grains of clastic origin (Plate 2, figs. 3, 5, and 6). Such a texture suggests a diagenetic or metamorphic origin for these minerals.

Many of the subgraywacke units are graded from coarse at the bottom to fine at the top; others are not graded, yet poorly sorted. They range in thickness from a few millimeters to, rarely, one meter. The thinner units (less than a few centimeters), because of lack of exposure, faulting, or tapering out, could only be traced laterally for distances of from 3 to 10 m; thicker units, however, could occasionally be traced up to 75 m. The thicker graded units (>10 cm) commonly contain particles whose maximum sizes range from granule in their lower parts to silt and clay in their upper parts (Plate 1). Thinner units rarely contain any particles >1 mm in diameter. Thin-section size distribution analyses of two thin, graded subgraywacke units from the Big Cottonwood Canyon area are shown as reconstructed weight frequency cumulative curves (after Greenman, 1951) in Plate 3 with accompanying photomicrographs. The size analyses indicate that grading in these units is far from regular. Abrupt transitions in grain size can be seen in the photomicrograph of BC–44 and in the size distribution curves of BC–6. Bent and twisted fine-grained phyllitic subgraywacke fragments occur in the middle and upper portions of graded units and throughout the entire thickness of non-graded poorly sorted units. Plate 1 shows a typical graded subgraywacke bed containing one long, twisted subgraywacke fragment (opposite the 11 cm mark) and many smaller fragments. In the thin sections of all subgraywackes observed by the author, many small, bent and twisted subgraywacke fragments occur among the clastic sand grains (Plate 2, fig. 3). Such fragments, which range from 0.5 mm to 10 cm in length, are particularly abundant in subgraywacke beds >10 cm thick.

Conglomeratic subgraywacke, which volumetrically constitutes a very small part of the total late Precambrian section (≤10 percent), is an important rock type for interpreting the environment of sedimentation and perhaps the associated climatic conditions at the time of sedimentation. These rocks have previously been mapped and interpreted as tillites (Blackwelder, 1932, p. 297; Crittenden and others, 1952, p. 4–6).

The best exposures of conglomeratic subgraywackes (and associated subgraywackes) are found in the central and northern Wasatch Mountains, on Fremont Island and Little Mountain in the Great Salt Lake area, in the southern Sheeprock Mountains, and in the southern Deep Creek Range (all located in northwestern Utah). Although a relatively thick section of conglomeratic subgraywacke is found in the northern Bannock Range in southern Idaho, it is for the most part poorly exposed.

The conglomeratic subgraywackes are confined to a belt ranging from 80 to 125 km wide which extends south from lat. 43° N. to 40° N. and then southwest to approximately long. 114° 30′ W.; the total length of the belt is about 800 km. Conglomeratic subgraywacke was not found west of the Snake Range in eastern Nevada.

The conglomeratic subgraywackes are usually black to greenish-black in color and contain from 20 to 40 percent of clasts ≥3 cm in diameter. A large majority of the clasts are from 3 to 15 cm in diameter and a few were found >2 m in diameter. Typically the clasts ≥3 cm in size do not touch one another but are distributed randomly in the dark matrix. In a very small proportion of these rocks (<1 percent), the clasts exceed the matrix in volume and touch one another thus forming true conglomerate. The dark matrix, which is commonly quartzitic, is rather arbitrarily defined in this investigation as all particles ≤2 mm in size. Using this definition, it constitutes from 60 to 80 percent by volume of the rocks. Although a complete gradation exists between matrix and rock clasts, field and thin-section studies suggest 2 mm as a convenient dividing point between the two. The matrix, in turn, can be described as a coarse fraction (particles from 0.1 to 2 mm) and a fine fraction (particles <0.1 mm). The fine fraction is in many cases mineralogically similar to the matrices of associated subgraywackes. Some of the conglomeratic subgraywacke matrices are highly quartzitic and could perhaps more appropriately be thought of as conglomeratic quartzites. However, because of the arbitrary division between these two rock types, conglomeratic subgraywacke, unless otherwise specified, will be used to include also conglomeratic quartzite.

Individual conglomeratic subgraywacke units, which commonly grade both vertically and horizontally into subgraywacke, have elongate lenticular shapes and usually range from 1 to 7 m in thickness (averaging about 3 m). Some of them could be traced horizontally for distances up to 75 m before they tapered out (grading into subgraywacke) or were covered by Recent or Pleistocene deposits. The intimate intertonguing and gradational nature of conglomeratic subgraywackes, subgraywackes, and quartzites occur on scales ranging from a few centimeters (Plate 2, figs. 7 and 8) to a few meters (Fig. 5A) or several tens of meters (Fig. 5B).

Unlike most nonreworked Pleistocene and Recent glacial tills which contain a large majority of broken, very angular particles in the sand size range, the late Precambrian conglomeratic subgraywackes contain a large majority of subrounded to well-rounded particles in this size range (Plate 2, fig. 2). The larger clasts in the conglomeratic subgraywackes, which are dominantly plutonic and quartzitic rock fragments, also have rounded to well-rounded shapes (Fig. 5C; Plate 2, figs. 5, 7, and 8). Wedge- and tabular-shaped clasts, which are common in many Pleistocene and Recent tills (Wentworth, 1936), were not found in the conglomeratic subgraywackes. A few plutonic and quartzite boulders were found, however, with one or more semiflat surfaces (facets?). Striated clasts, which are minor yet characteristic features of most Recent and Pleistocene tills, were not found by the author in any of the

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PLATE 2. TEXTURES AND STRUCTURES OF SUBGRAYWACKES AND CONGLOMERATIC SUBGRAYWACKES FROM NORTHERN UTAH

Figure 1. Bent and twisted subgraywacke fragments in a typical conglomeratic subgraywacke unit from Big Cottonwood Canyon

Figure 2. Subgraywacke fragment in the matrix of a typical conglomeratic subgraywacke from Big Cottonwood Canyon (unx nicols)

Figure 3. Twisted shaly subgraywacke fragment in a graded subgraywacke bed from Big Cottonwood Canyon (unx nicols)

Figure 4. Augen-shaped sand grains in a typical subgraywacke metamorphosed to the greenschist facies from Fremont Island (unx nicols)

Figure 5. Subgraywacke from Big Cottonwood Canyon showing well rounded quartz sand grains, most of which are embayed by matrix minerals (unx nicols)

Figure 6. Subgraywacke from Big Cottonwood Canyon showing clastic sand grains embayed by matrix chlorite, biotite, and carbonaceous matter (unx nicols)

Figure 7. Interfingering and gradational subgraywacke (left) and conglomeratic subgraywacke (upper and right) at Dutch Peak in the Sheeprock Mountains

Figure 8. Intimately interbedded subgraywacke and conglomeratic subgraywacke from the Sheeprock Mountains

Figure 1

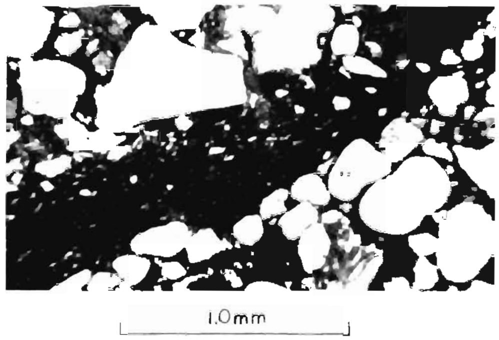



Figure 2

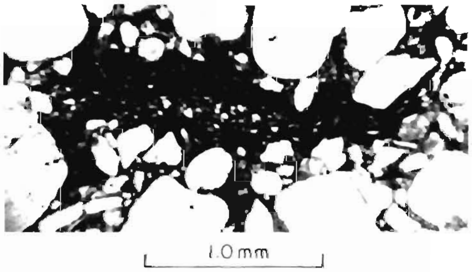



Figure 3

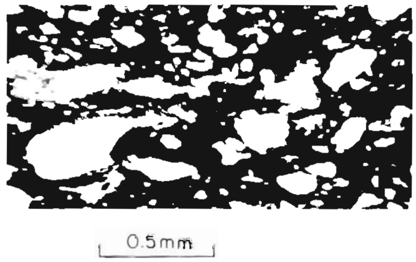



Figure 4

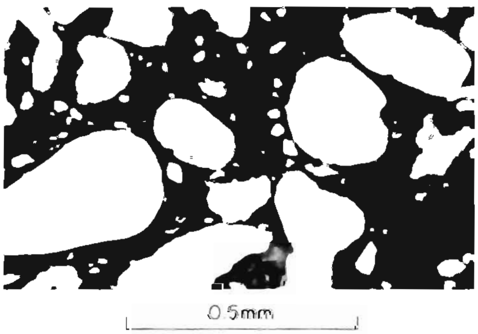



Figure 5

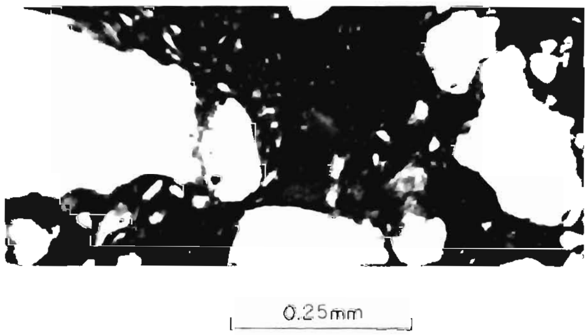



Figure 6

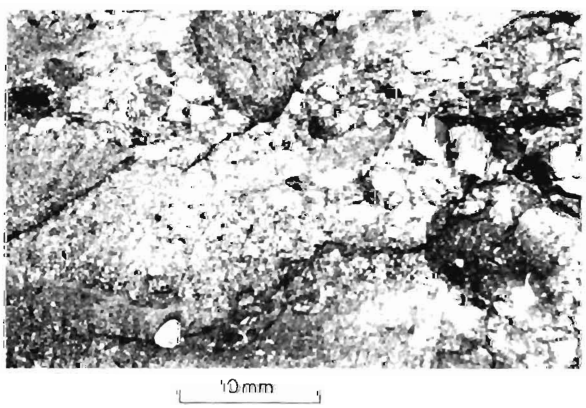
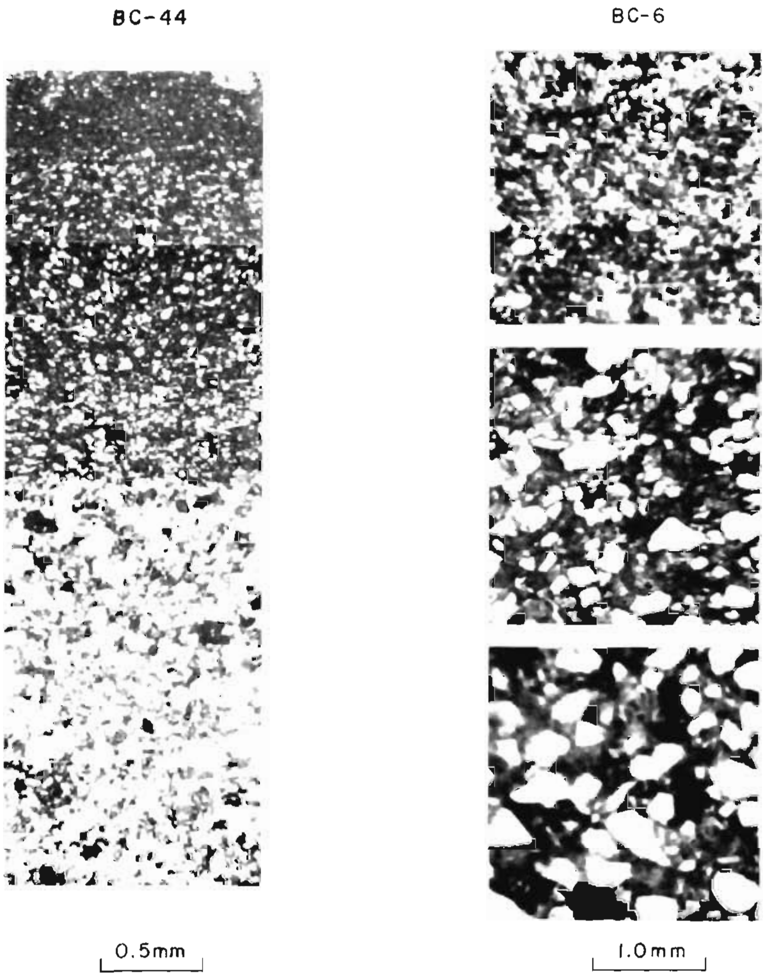



Figure 7

Figure 8

TEXTURES AND STRUCTURES OF SUBGRAYWACKES AND CONGLOMERATIC SUBGRAYWACKES FROM NORTHERN UTAH



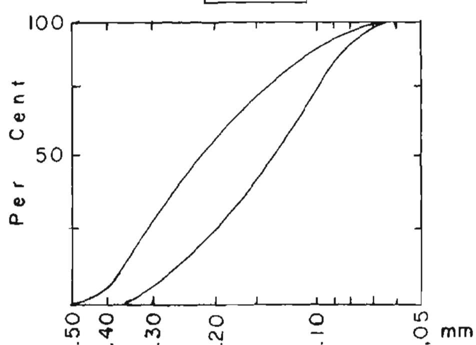




Figure 1

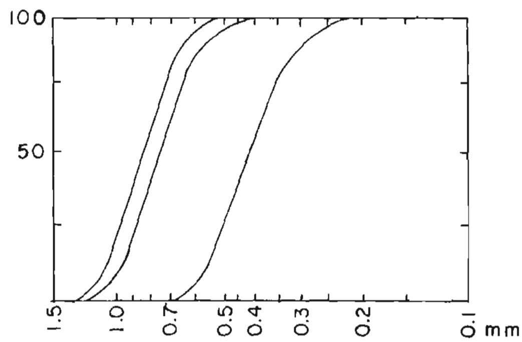



Figure 2

GRADED SUBGRAYWACKES BC-6 AND BC-44
FROM THE BIG COTTONWOOD CANYON AREA

conglomeratic subgraywackes in northern Utah. Blackwelder (1932), however, has reported both faceted and striated clasts from the Little Mountain locality. The presence of a few faceted (and perhaps striated) clasts in the conglomeratic subgraywackes led Blackwelder and others to interpret these rocks as tillites. However, the fact that wind, surf, running water, and terrestrial mudflows have all been reported to facet and striate clasts indicates that the presence of these features is not necessarily diagnostic of glacial deposition or of glacial climatic conditions.

Bent and twisted subgraywacke fragments ranging from a few millimeters to nearly 10 m in length and from $\leq 1$ mm to 5 cm thick are found in varying proportions in all of the conglomeratic subgraywackes. In Mineral Fork and Mill B South Fork in the Big Cottonwood Canyon area (Fig. 2), they constitute up to 50 per cent of some conglomeratic subgraywacke units. The fact that these fragments are twisted and intimately interwoven among the other clasts (Fig. 5C; Plate 2, figs. 1–3) suggests that they were still soft and plastic when eroded and deposited. These contorted and twisted beds in the conglomeratic subgraywackes are suggestive of slumping.

### *Mineralogy*

*Quartz.* Quartz is the most abundant mineral in most of the subgraywackes and conglomeratic subgraywackes (Tables 3 and 4). It occurs as rounded sand grains, as fine, more angular shaped grains in the rock matrices, and as a major constituent of rock fragments. In thin section, almost all of the quartz grains show undulatory extinction in varying degrees and commonly they are dusty with black inclusions. Apatite and rutile occur in some of the larger quartz grains and in the quartz contained in rock fragments. All of the obviously clastic grains are embayed in varying degrees by diagenetic and metamorphic matrix minerals. The more quartzitic subgraywackes and conglomeratic subgraywackes are composed almost entirely of quartz grains and quartzite rock fragments of various sizes. Although a large amount of the fine-grained quartz could not be distinguished from plagioclase, both chemical and X-ray diffraction data suggest that it composes a very large fraction of the total quartz + plagioclase mode.

*Plagioclase.* In the unmetamorphosed subgraywackes and conglomeratic subgraywackes plagioclase occurs as subrounded to rounded sand grains of clastic origin, as a fine diagenetic matrix constituent, and as a major mineral in plutonic rock fragments. Obvious clastic plagioclase was not found in any samples metamorphosed to the greenschist facies. Rather, the fact that most of the plagioclase in the metamorphosed subgraywackes and conglomeratic subgraywackes is euhedral, suggests that it is of metamorphic origin. The plagioclase, particularly in the metamorphic rocks, is altered in varying degrees to calcite and sericite which in some cases almost completely replace the plagioclase crystals. Zoning was not found in any of the plagioclase in these rocks. As with quartz, plagioclase, whether of metamorphic or clastic origin, is embayed in varying degrees by diagenetic or metamorphic matrix constituents.

Plagioclase compositions were determined by measuring the $\alpha$ refractive index of two to five grains per sample in polarized Na light using standard immersion liquids. Corresponding An contents were calculated using the regression equations of Chayes (1952) and the results are given in Table 5. It is apparent from the table that calcic oligoclase is the dominant plagioclase in both the unmetamorphosed and metamorphosed samples. The variation in plagioclase composition is more restricted, however, in the subgraywackes, conglomeratic subgraywackes, and plutonic rock fragments metamorphosed to the greenschist facies. This extreme uniformity of composition, together with the occurrence of plagioclase as euhedral crystals, suggests that plagioclase in the meta-

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PLATE 3. GRADED SUBGRAYWACKES BC–6 AND BC–44 FROM THE BIG COTTONWOOD CANYON AREA

Figure 1. Photomicrograph and reconstructed weight frequency (percent) size distribution curves of subgraywacke bed BC–44, 3.75 mm thick (unx nicols), (unx-uncrossed)

Figure 2. Photomicrographs and reconstructed weight frequency (percent) size distribution curves of subgraywacke bed BC–6, 2.5 cm thick. Photomicrographs are of the lower (L), middle (M), and upper (U) portions of the bed. (unx nicols)

TABLE 5. SUBGRAYWACKE AND CONGLOMERATIC SUBGRAYWACKE PLAGIOCLASE COMPOSITIONS (PERCENT AN)

| | x̄* | M.D. |
|---|---|---|
| **Unmetamorphosed samples** | | |
| *Subgraywackes* | | |
| BC-32 | 41 | 7 |
| BC-44 | 25 | 4 |
| BC-6 | 28 | 10 |
| BC-45 | 50 | 4 |
| *Conglomeratic subgraywackes* | | |
| GYU-4 | 33 | 2 |
| BC-4 | 43 | 5 |
| BC-21 | 35 | 2 |
| BC-38 | 27 | 4 |
| *Plutonic boulders* | | |
| BC-23 | 30 | 0 |
| BC-35 | 33 | 0 |
| BC-36 | 28 | 0 |
| BC-37 | 33 | <1 |
| x̄ | 34 | |
| Range | 29–50 | |
| **Metamorphosed samples** | x̄ | M.D. |
| *Subgraywackes* | | |
| F-3 | 33 | 1 |
| F-54 | 28 | <1 |
| *Conglomeratic subgraywackes* | | |
| LM-5 | 31 | 1 |
| P-8 | 31 | 1 |
| F-70 | 30 | 1 |
| F-71 | 34 | <1 |
| LM-23 | 33 | <1 |
| *Plutonic boulders* | | |
| F-77 | 31 | <1 |
| F-78 | 31 | 0 |
| F-80 | 34 | 0 |
| F-104 | 33 | 0 |
| LM-9 | 33 | <1 |
| LM-10 | 31 | <1 |
| LM-14a | 33 | 0 |
| LM-27 | 33 | <1 |
| x̄ | 32 | |
| Range | 28–34 | |

* Mean values (x̄) and mean deviations (M.D.) of five or more grains.

morphic rocks is of metamorphic rather than clastic origin. Diagenesis has not greatly affected the plagioclase in the unmetamorphosed rocks (especially the subgraywackes) as evidenced by both the comparatively wide range of plagioclase composition and the rounded (clastic) shapes of many of the larger grains. However, the fine matrix plagioclase in the unmetamorphosed rocks is euhedral and appears to be of diagenetic origin.

Although the abundance of plutonic rock fragments in the subgraywackes and conglomeratic subgraywackes suggests that most of the clastic plagioclase in these rocks was derived from similar sources, X-ray diffraction studies of the unmetamorphosed rocks were undertaken in an attempt to determine if any volcanic (high-temperature) plagioclase was present. Differences between the $(1\bar{3}1)$ and $(131)$ plagioclase diffraction peaks were carefully measured. Using the data of Smith and Yoder (1956), the maximum $\Delta$ $(131-1\bar{3}1)$ values indicate that the plagioclase in these rocks is the low-temperature variety.

*Microcline.* Microcline is an extremely minor component in the subgraywackes and conglomeratic subgraywackes. When found, it occurs as isolated clastic grains (as evidenced by its subrounded to rounded shapes) or in plutonic rock fragments. X-ray diffraction studies indicate that microcline is negligible or absent in the fine-grained matrices. Like quartz and plagioclase, microcline grains are strongly embayed by surrounding matrix minerals and are sericitized in varying degrees. Grid twinning, however, is commonly preserved even in the highly sericitized grains. With very few exceptions, microcline occurs only in the unmetamorphosed rocks (Tables 3 and 4). The rare microcline found in the metamorphosed samples is highly sericitized and commonly surrounded by what appears to be a metamorphic reaction rim of muscovite.

*Muscovite and paragonite.* Muscovite is a major constituent of the fine-grained matrices of most of the unmetamorphosed and low-grade metamorphic subgraywackes and conglomeratic subgraywackes. It also occurs ubiquitously as a feldspar alteration product (sericite). Paragonite, which was identified by X-ray diffraction, occurs intimately mixed with muscovite in many of the subgraywackes metamorphosed to the greenschist facies. The fact that both muscovite and paragonite commonly penetrate and embay clastic sand grains indicates a diagenetic or metamorphic origin for these minerals.

In order to investigate the compositions of muscovite and paragonite in the metamorphosed subgraywackes and conglomeratic subgraywackes, the (002) and (005) reflections in total rock samples in which biotite was negligible or absent were measured on a Norelco diffractometer using the (101) and (112) quartz reflections as built-in internal standards. The results are shown in Table 6 with corresponding paragonite mole fractions ($N_{Pa}$) which were calculated from the regression equation of Zen and Albee (1964, p. 915) and the (002) spacings. It is obvious from the tables that only very small amounts of Na and K are contained in the muscovite and paragonite, respectively. Small ranges in Na and K solid solution in these minerals are in agreement with the findings of Zen and Albee (1964) which suggest that only limited solid solution occurs in muscovite and paragonite found in low-grade metamorphic terranes.

TABLE 6. (002) AND (005) REFLECTIONS AND $N_{Pa}$* VALUES OF MUSCOVITE AND PARAGONITE FROM METAMORPHOSED SUBGRAYWACKES AND CONGLOMERATIC SUBGRAYWACKES

| | (002) | | (005) | | |
|---|---|---|---|---|---|
| | 2θ(CuKα) | d(Å)† | 2θ(CuKα) | d(Å)† | $N_{Pa}$ |
| F-3 | 8.81 | 10.028 | 45.25 | 2.002 | 0.01 |
| F-7 | 8.82 | 10.017 | 45.16 | 2.006 | 0.04 |
| F-14 | 8.83 | 10.006 | 45.27 | 2.001 | 0.07 |
| F-24 | 8.81 | 10.028 | 45.24 | 2.003 | 0.01 |
| F-54 | 8.82 | 10.017 | 45.34 | 1.998 | 0.04 |
| F-59 | 8.82 | 10.017 | 45.36 | 1.998 | 0.04 |
| F-70 | 8.84 | 9.995 | 45.37 | 1.997 | 0.09 |
| F-71 | 8.85 | 9.983 | 45.40 | 1.996 | 0.12 |
| LM-23 | 8.81 | 10.028 | 45.43 | 1.995 | 0.01 |
| PP-26 | 8.81 | 10.028 | 45.37 | 1.997 | 0.01 |
| PP-29 | 8.82 | 10.017 | 45.35 | 1.998 | 0.04 |
| F-3 | 9.15 | 9.657 | .. | .. | 0.88 |
| F-7 | 9.19 | 9.625 | .. | .. | 0.95 |
| F-14 | 9.16 | 9.646 | .. | .. | 0.91 |

* $N_{Pa}$–mole fraction of paragonite
† Mean values of two or more runs with mean deviation ≤ 0.001

*Biotite.* Biotite is a minor mineral in many of the subgraywackes and conglomeratic subgraywackes; however, in a few of the unmetamorphosed samples it is an important constituent (for example, GYU-4 and BC-21). In these samples it occurs as fine-grained crystals intergrown with muscovite, chlorite, and carbonates. The fact that it commonly embays clastic sand grains suggests a diagenetic origin. The biotite found in all of the samples metamorphosed to the greenschist facies is partially to entirely converted to chlorite, suggesting that it was not stable during low-grade metamorphism.

*Chlorite.* Chlorite occurs as a fine matrix constituent in most subgraywackes and conglomeratic subgraywackes. The fact that it commonly embays clastic grains suggests a diagenetic or metamorphic origin for most of it. In the unmetamorphosed samples it occurs intimately intergrown with quartz, biotite, muscovite, and carbonates. Occasional large patches of chlorite in both unmetamorphosed and metamorphosed samples appear to have originated by diagenetic or metamorphic recrystallization of biotite (and perhaps hornblende) as suggested by relict crystal outlines.

The Fe + Ti + Mn + Cr/Fe + Ti + Mn + Cr + Mg ratios (hereafter abbreviated f/fm) in the chlorites were determined using the X-ray diffraction method developed by Schoen (1962). The intensities of the (001), (002), and (003) basal spacings were measured on diffractograms and corresponding structure factors were calculated. The f values (f = total number of Fe + Mn + Cr + Ti atoms per 12 octahedral sites) and f/fm ratios were determined using the ratios of calculated structure factors to theoretical structure factors (Schoen, 1962). The results, given in Table 7, indicate that the chlorites from the greenschist-facies terrane are more Mg-rich than those from unmetamorphosed terrane.

*Carbonate Minerals.* Three carbonate minerals, calcite, siderite, and dolomite, occur in the subgraywackes and conglomeratic subgraywackes. The occurrence of each mineral is quite distinct. Calcite occurs as irregular patches of cementing material between clastic grains and as an alteration product of plagioclase. The fact that calcite is rare or absent in most unmetamorphosed samples, yet is quite common in some of the samples metamorphosed to the greenschist facies, suggests a metamorphic origin for much of it.

Siderite, on the other hand, was found only in the unmetamorphosed subgraywackes (Table 2). It occurs in these rocks as a fine-grained matrix component intimately intergrown with chlorite and biotite. The fact that it strongly embays clastic quartz and feldspar grains indicates a diagenetic origin. Commonly it is concentrated in bands from 0.5 to 1 mm thick in the subgraywackes: these bands are usually more abundant in the chlorite- and biotite-rich subgraywackes and commonly occur in the upper portions of graded units. Siderite becomes an exceedingly abundant matrix component locally as shown in sample BC–3.

Dolomite occurs only as rock fragments and rare oölites. Most fragments are from 0.5 to 2 mm in size but in some places they attain diameters of 30 to 50 cm. Like the quartz grains they are well rounded and commonly embayed by matrix minerals.

*Chloritoid.* Although a minor amount of chloritoid occurs in the subgraywackes near the base of the Big Cottonwood Canyon section, it is limited for the most part to subgraywackes metamorphosed to the greenschist facies. It occurs as euhedral porphyroblasts ranging from ~0.1 mm to over 2 cm in length and it com-

TABLE 7. STRUCTURE FACTORS AND F/FM RATIOS OF CHLORITES FROM SUBGRAYWACKES AND CONGLOMERATIC SUBGRAYWACKES*

| | $F_{(001)}$ | $F_{(002)}$ | $F_{(003)}$ | f† | f/fm‡ | | |
|---|---|---|---|---|---|---|---|
| | | | | | | $\bar{x}$ | M.D. |
| | | | Unmetamorphosed samples | | | | |
| BC–2 | .080 | .366 | .210 | 9.9 | .83 | | |
| BC–10b | .082 | .392 | .271 | 7.4 | .62 | | |
| BC–44 | .064 | .249 | .167 | 7.3 | .61 | | |
| BC–4 | .063 | .265 | .166 | 8.3 | .69 | | |
| BC–21 | .061 | .246 | .166 | 7.3 | .61 | .68 | .06 |
| BC–38 | .090 | .338 | .204 | 8.3 | .69 | | |
| BC–39 | .062 | .266 | .167 | 8.5 | .71 | | |
| GYU–4 | .074 | .260 | .167 | 7.4 | .62 | | |
| BC–20b | .107 | .352 | .323 | 8.9 | .74 | | |
| | | | Metamorphosed samples | | | | |
| F–2 | .138 | .548 | .395 | 6.8 | .57 | | |
| F–3 | .139 | .555 | .387 | 7.1 | .59 | | |
| F–7 | .173 | .431 | .307 | 5.0 | .42 | | |
| F–14 | .088 | .476 | .348 | 7.2 | .60 | | |
| F–20 | .161 | .640 | .448 | 6.9 | .58 | | |
| F–24 | .065 | .330 | .277 | 5.7 | .48 | | |
| F–54 | .116 | .458 | .382 | 6.6 | .55 | | |
| F–59 | .106 | .453 | .340 | 6.6 | .55 | | |
| PP–29 | .126 | .413 | .258 | 7.4 | .62 | .49 | .08 |
| PP–26 | .109 | .301 | .226 | 4.9 | .41 | | |
| P–8 | .078 | .339 | .263 | 6.1 | .51 | | |
| F–55 | .111 | .457 | .387 | 5.1 | .43 | | |
| F–71 | .081 | .338 | .281 | 5.3 | .44 | | |
| LM–5 | .073 | .271 | .241 | 5.0 | .42 | | |
| LM–23 | .070 | .268 | .197 | 6.3 | .53 | | |
| RC–3 | .062 | .215 | .204 | 3.7 | .31 | | |
| RC–4 | .065 | .237 | .232 | 3.5 | .29 | | |
| Sh–3 | .117 | .418 | .344 | 5.1 | .43 | | |

* Structure factors represent mean values of two or more runs; corresponding f/fm ratios have mean deviations $\leq 0.008$

† f–number of Fe + Mn + Cr + Ti atoms per 12 octahedral sites

‡ $f/fm - \frac{Fe + Mn + Cr + Ti}{Fe + Mn + Cr + Ti + Mg}$ ratio

monly shows well developed hourglass structure. Most of the chloritoid is aligned with its long axis from 45° to 90° with the foliation direction. "Shadows" of quartz and muscovite occur commonly on the sides of the near-vertical chloritoid crystals. Also, most of the chloritoid crystals contain an abundance of chlorite, quartz, and muscovite inclusions around which they have grown during metamorphism. They are commonly concentrated in the upper portions of graded subgraywacke units where there are relatively large amounts of biotite and chlorite from which they have probably grown.

Chloritoid concentrates from some of the subgraywackes were scanned on a Norelco X-ray diffractometer to determine the crystal system of the chloritoid using the data of Halferdahl (1961, p. 82). The (101) quartz peak was used as an internal standard. The fact that strong $(1\bar{1}2)$ reflections occurred in all of the samples analyzed, together with the absence of diagnostic monoclinic reflections, indicates that the chloritoid found in the subgraywackes is triclinic.

*Other Minerals.* Because it was often difficult to distinguish ilmenite and magnetite from carbonaceous matter (including graphite), these opaque substances were combined in the modal analyses (Tables 3 and 4). Polished-section studies, however, indicate that ilmenite and, to a lesser degree, magnetite compose a large percentage of this fraction. They occur as irregular patches in the rock matrices engulfing and embaying clastic grains in some places. Carbonaceous matter (including graphite), when positively identified, occurs as very fine particles evenly distributed in the rock matrices. Hematite and limonite occur as irregular patches, crack fillings, and pseudomorphs of other iron oxides and sulfides. Such an occurrence suggests a secondary origin for these minerals. Some hematite in the metamorphosed rocks, however, occurs as discrete euhedral grains in the matrices suggesting a metamorphic origin.

Acicular rutile crystals and prismatic apatite crystals occur as small inclusions in the larger clastic quartz grains in most of the subgraywackes and conglomeratic subgraywackes. Zircon usually occurs as subrounded to rounded crystals. Clastic crystals of sphene also were found in a few of the conglomeratic subgraywackes. Pyrite or pseudomorphs of limonite after pyrite is a common metamorphic mineral in the subgraywackes metamorphosed to the greenschist facies. Its metamorphic origin is attested to by its well developed crystal faces. Minor amounts of very fine-grained epidote and actinolite, both of metamorphic origin, were found in conglomeratic subgraywacke samples P-8 and Sh-3.

### *Rock Fragments*

Rock fragments which constitute from 10 to 60 percent of the conglomeratic subgraywackes and from 1 to 10 percent of the subgraywackes, range in size from <1 mm to over 2m. Of the small rock fragments, quartzite and reworked subgraywacke are the most abundant; quartzite and plutonic rocks predominate in the larger clasts. However, because there is a complete gradation between mineral grains and rock fragments in the size range 0.5 to 2 mm, rock fragments in this size range (0.5–2 mm) are not distinguished as such in the modal analyses but their constituent minerals are counted together with the other minerals in the rocks.

Quantitative estimates of the lithology of rock fragments > 2 mm in size from conglomeratic subgraywackes were determined by counting and recording the corresponding lithology of 2500 to 5000 points (at 30 cm intervals) in individual conglomeratic subgraywacke units (Table 8). This macromodal analysis indicates that plutonic and quartzite fragments are the most abundant lithologies. However, units occur locally that are rich in reworked subgraywacke (Table 8, no. 7) and dolomite (Table 8, no. 3). The carbonate rock fragments consist chiefly of fine-grained yet well crystallized dolomite. Metadiabase rock fragments are uncommon in most of the subgraywackes and conglomeratic subgraywackes, but attain abundances up to 10 percent in some units in the northern Bannock Range. Chert is is a minor yet ubiquitous rock fragment in all of the conglomeratic subgraywackes.

The plutonic rock fragments, which in this investigation include granite, pegmatite, and gneiss, are very similar petrologically to the underlying and adjacent Precambrian plutonic-metamorphic terranes that probably served as source areas. Abnormal amounts of chlorite and muscovite in some of these rocks result from diagenetic recrystallization of feldspars and biotite.

### *Geochemistry*

The chemical composition of the subgraywackes and conglomeratic subgraywackes is highly variable as indicated by the large mean deviations of most major and trace element concentrations given in Tables 3 and 4. These rocks are, however, lower in CaO and higher in $K_2O$, $Al_2O_3$, $TiO_2$, and Rb than typical subgraywackes and graywackes tabulated in Pettijohn (1963). The large amounts of K and Al, the low amounts of Ca, and the high quartz/feldspar ratios indicate a high degree of maturity for the original sediments. $SiO_2$, MnO, total Fe, Sr, Zr, and Ni are highly variable in these rocks and overlap most subgraywacke and graywacke averages (Pettijohn, 1963).

and limonite. The limestone unit in the northern Bannock Range is composed chiefly of coarse twinned calcite with the same accessory minerals as in the dolomites. X-ray diffraction measurements of the (104) carbonate spacings indicate that the dolomite and calcite are nearly stoichiometric containing $\leq 0.1$ percent excess Ca or Mg respectively (using the data of Goldsmith and Graf, 1958).

## ENVIRONMENT OF SEDIMENTATION

At least two different sedimentary environments existed in the northeastern Great Basin during the late Precambrian. The first and most extensive gave rise to the massive, commonly cross-bedded quartzites (and carbonate rocks) which constitute most of the section. Their gross similarity to overlying Paleozoic quartzites suggests a similar environment of sedimentation—a widespread shallow marine environment. The presence of mud cracks in some of these quartzites suggests that near shore, perhaps tidal flat conditions existed periodically.

The other major sedimentary environment, of much smaller extent, gave rise to the subgraywackes and conglomeratic subgraywackes. Although textures and structures found in these rocks are somewhat ambiguous, the author tentatively postulates turbidity current and subaqueous mudflow origins, respectively. The occurrence of poorly sorted, commonly graded beds; bent and twisted apparently reworked subgraywacke fragments; and contorted subgraywacke beds suggests such origins. The fact that the conglomeratic subgraywackes commonly grade laterally into graded subgraywacke beds suggests that many of the postu-

TABLE 8. LITHOLOGY OF ROCK FRAGMENTS IN CONGLOMERATIC SUBGRAYWACKES (VOLUME PERCENT)

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 |
|---|---|---|---|---|---|---|---|---|
| Quartzite | 13 | 15 | 10 | 20 | 10 | 21 | 11 | 10 |
| Plutonic | 17 | 14 | 1 | 20 | 25 | 19 | 8 | 16 |
| Carbonate | tr* | tr | 11 | tr | tr | 1 | tr | 1 |
| Subgraywacke | tr | tr | tr | .. | tr | 3 | 38 | 1 |
| Matrix† | 70 | 71 | 78 | 60 | 65 | 56 | 43 | 72 |

* tr–trace (<1 percent)
† particles ≤ 2 mm
1 North end of Fremont Island
2 North end of Fremont Island
3 Rock Canyon area, east of Utah Lake
4 Dutch Peak, Sheeprock Mountains
5 Northern end of Antelope Island
6 Mineral Fork, Big Cottonwood Canyon area
7 Mineral Fork, Big Cottonwood Canyon area
8 Mineral Fork, Big Cottonwood Canyon area

## CARBONATE ROCKS

Carbonate rocks are extremely rare in the late Precambrian section. A few thin dolomite beds were found interbedded with quartzite and subgraywacke in the southern Promontory Range (Fig. 4), on Fremont Island (Fig. 3), on the northern end of Antelope Island, in the Canyon Mountains, and in the Southern Deep Creek Range (dolomite marble), all in northwestern Utah. These units, which are a few centimeters to 2 m thick and very massive, are typically devoid of bedding and other primary structures. Under the microscope the carbonate minerals appear to be entirely recrystallized. One limestone unit 40 m thick occurs in the section in the northern Bannock Range; however, field relationships indicate that it may be Paleozoic in age.

The mineralogy of the carbonate rocks is very simple compared to the clastic rocks. The dolomites are composed of very fine-grained dolomite (from 98 to 100 percent as determined optically and by X-ray diffraction) with occasional grains of quartz, plagioclase, muscovite, lated mudflows, perhaps originating by subaqueous slumping, gave rise to turbidity currents which deposited associated subgraywackes. The fact that gradational relationships exist between these rocks on various scales suggests that the postulated subaquaous mudflows and turbidity currents ranged in size and lateral extent by many orders of magnitude.

Because subgraywackes associated with conglomeratic subgraywackes (presumed tillites) are often interpreted as glacial varves, it is of interest to compare the late Precambrian subgraywackes of northern Utah with Pleistocene varves. Distal varves, which commonly range from 1 to 5 cm thick, contain only a few percent of particles >0.2 mm in size (Sauramo, 1923; Arrhenius, 1947; Antevs, 1951) and in this way differ strikingly from the late Precambrian subgraywacke beds of the same thickness which contain large amounts of particles between 0.2 and 1 mm in size. Although proximal varves, which are usually 0.5 to 2 m thick and contain an abundance of particles in the sand- to pebble-size range, may have size distributions similar to some of the Precambrian subgraywacke beds, they differ from the subgraywackes in that they are closely associated with oscenters or kame terraces (DeGeer, 1940). Both distal and proximal varves differ from the subgraywacke beds in that carbonaceous matter, if present, is concentrated in the upper winter laminae, whereas it is rather evenly distributed or concentrated in the middle and lower portions of the subgraywacke beds. These textural and structural differences suggest that the sediment sources, mechanics of deposition, and environment of sedimentation, or all three, may have been quite different during the formation of the late Precambrian subgraywackes than during the formation of glacial varves.

If a turbidity current and subaqueous mudflow origin is tentatively accepted for the late Precambrian subgraywackes and conglomeratic subgraywackes of northern Utah, any model of erosion prior to redeposition by these agents must account for the abundance of well rounded clasts ranging in size from sand to boulder and the deposition of these clasts with finer material at the sites of subaqueous slumping. Although boulders can become well rounded in a few kilometers by stream erosion, sand grains, unless they have survived many erosion cycles, are not appreciably rounded by agents other than wind and surf (Kuenen, 1959). Because the quartzite rock fragments in the subgraywackes and conglomeratic subgraywackes contain an abundance of angular to subrounded grains, it is therefore unlikely that the quartz sand derived from the quartzite source areas (from which the quartzite rock fragments originated) was appreciably rounded by fluvial processes. For this reason it is probable that wind or surf or both played a role in rounding sand grains in the subgraywackes. Among the known agents capable of rounding and, occasionally, faceting and striating boulders and then depositing them in a heterogeneous mixture with finer particles are: ice, subaqueous and terrestrial mudflows, and streams which vary greatly in competency such as glacial streams and streams of semi-arid and arid regions. One or some combination of these agents was probably responsible for depositing such material at the sites of proposed subaqueous slumping and turbidity current generation in northern Utah during the late Precambrian.

The large compositional variation between subgraywacke units, from those consisting almost entirely of quartz to those consisting almost entirely of mica and chlorite, may indicate corresponding variations in the compositions of the sediments being supplied to the sites of subaqueous slumping. Such variation may reflect slumping from different sites on a deltaic slope with quartz-rich turbidites (and mudflows) originating at near-shore sites and mica- and chlorite-rich ones at far-shore sites. On the other hand, if the sedimentary basins were partially enclosed (as in bays or lagoons), turbidites of one composition may have originated on one side, while those of another composition originated on another side and hence the compositional variation may reflect two or more source areas of differing composition.

Recent investigations of presumed tillites from Massachusetts (Dott, 1961), the Western Congo (Schermerhorn and Stanton, 1963), and western France (Winterer, 1965) report textures and structures similar to those reported in this investigation, suggesting that they were deposited as subaqueous mudflows or slides. A subaqueous mudflow origin does not, however, necessarily eliminate the possibility of contemporaneous glaciation. The occasional faceted and striated clasts found in these rocks, if of glacial origin, may have been derived from reworking and redeposition of glacial till deposited by contemporaneous continental ice sheets. However, the possibility also exists that some or all of these clasts were derived from tills of contemporaneous alpine (or piedmont) glaciers and hence, reflect only local or regional but not continental climatic conditions. The problem of reworking is even more complicated by the fact that clasts from older tillites may also have been reworked and redeposited in subaqueous mudflows.

The abundant well rounded sand grains in the late Precambrian subgraywackes and conglomeratic subgraywackes of the northeastern Great Basin, which are not reported in glacial varves or in most nonreworked Pleistocene tills, also do not necessarily eliminate the possibility of penecontemporaneous glaciation. Such grains may have been derived from dune, beach, or multicycled fluvial sands which were mixed and redeposited with glacial material. Dunes, in fact, are quite common in Recent and Pleistocene glacial outwash plains (Flint, 1957). Although the subgraywacke beds in northern Utah may not be genetically related to glacial varves, the possibility still remains that they are composed in part or entirely of redeposited contemporaneous glacial, glacial-fluvial, or glacial-marine material.

The possibilities of reworking and redepositing mixtures of clasts derived both from contemporaneous sediments and from older rocks of either glacial or nonglacial origin, or both, greatly complicate the problem of interpreting the erosional history of the late Precambrian subgraywackes and conglomeratic subgraywackes from northern Utah. Until it is possible to identify redeposited contemporaneous glacial material and to distinguish textures and structures of known glacial origin from similar nonglacially produced textures and structures, care should be used in postulating pre-Pleistocene glacial epochs from the existence of conglomeratic subgraywackes or other pebbly mudstones.

## PROVENANCE

Plutonic- and quartzite-rich source areas are indicated for most of the late Precambrian sediments in northern Utah from the rock fragments they contain. The underlying and adjacent older Precambrian terranes in northern Utah and southwestern Wyoming, which contain an abundance of plutonic and gneissic rock types, represent probable sources for the plutonic clasts. Although quartzite is not a common rock type in the older Precambrian plutonic-metamorphic terranes, great thicknesses of quartzite underlie the subgraywackes and conglomeratic subgraywackes. The lateral equivalents of some of these quartzites may have served, in part or entirely, as sources for the quartzite clasts as well as for much of the quartz in these rocks.

The fact that amphibolite rock fragments were not found in any of the subgraywackes or conglomeratic subgraywackes is somewhat puzzling because amphibolite composes from 5 to 15 percent of the older Precambrian plutonic-metamorphic terranes. Its absence may be related, in part, to the relatively rapid weathering rates of hornblende and calcic plagioclase as indicated by the work of Goldich (1938, p. 48–53). It is possible that the calcic plagioclase (andesine) of clastic origin in some of the subgraywackes and conglomeratic subgraywackes was derived from amphibolites in the older Precambrian source areas.

The major provenance problem in the subgraywackes and conglomeratic subgraywackes is that of the nature and origin of the original matrix minerals. Although, with the exception of some quartz and perhaps feldspar, none of the original fine-grained matrix components of these rocks were positively identified, it is likely that the original sediments contained large amounts of sheet structure silicates and varying amounts of quartz and feldspar. The possible sources for such material include: pyroclastic material derived from contemporaneous volcanic eruptions, shaly sections in the source areas, and a weathered mantle on the plutonic source areas. The absence of volcanic rock units and clasts as well as high-temperature (volcanic) feldspar in the late Precambrian section does not support a volcanic origin. If the matrix material was derived primarily from the reworking of Precambrian shaly sections, it is necessary that such sections were either completely removed by subsequent erosion or that they are now concealed. Although it is not possible with the present data to eliminate this latter possibility, the third possible origin, that of a weathered mantle on the plutonic source areas, is tentatively favored by the author. Such weathered material may have been transported by either glacial ice and streams (together with rock flour) or by torrential streams and mudflows to the sites of subaqueous slumping.

## DIAGENESIS AND METAMORPHISM

All of the late Precambrian rocks in northern Utah possess textural, structural, and compositional evidence of diagenetic and metamorphic changes. The unmetamorphosed samples from the Big Cottonwood Canyon area appear to have undergone diagenetic recrystallization as evidenced by the penetration and embayment of clastic sand grains by chlorite and mica. The degree of embayment increases as does the average grain size with metamorphic grade. The clastic quartz sand grains in the quartzites and many of the subgraywackes and conglomeratic subgraywackes are commonly surrounded by quartz overgrowths of diagenetic or metamorphic origin. Such overgrowths are a common feature even in rocks metamorphosed to the greenschist facies. Most of the original sand grains and small rock fragments in the less quartzitic varieties of metamorphosed subgraywacke and conglomeratic subgraywacke are broken and stretched into augen shapes (Plate 2, fig. 4). Foliation is well developed in these rocks.

Inspection of the data in Tables 3 and 4 indicates that differences in mineralogy and in major and trace element concentrations occur between unmetamorphosed and metamorphosed terranes. Whether all such differences are real, however, is questionable because the number of samples analyzed is small compared to the variability between samples from a given terrane. The consistent mineralogical differences which appear to be related to progressive low-grade metamorphism include: increases in chloritoid, calcite, and pyrite, and decreases in microcline, siderite, and dolomite. The data in Tables 5 and 7 indicate that the An content of plagioclase and the f/fm chlorite composition also decrease between unmetamorphosed and metamorphosed terranes. The occurrence of siderite in some of the unmetamorphosed subgraywackes and conglomeratic subgraywackes suggests that a reducing environment may have existed during diagenesis.

The fact that the plagioclase in the low-grade metamorphic subgraywackes, conglomeratic subgraywackes, and plutonic rock fragments has a small compositional range ($An_{28-34}$) (Table 5) suggests that equilibrium was approached among the plagioclase crystals during regional metamorphism. Also, the fact that the diagenetic and low-grade metamorphic chlorites in the subgraywackes, and conglomeratic subgraywackes have distinct compositional ranges (Table 7) suggests that both diagenetic (high-Fe chlorite) and metamorphic (low-Fe chlorite) equilibria were approached in the chlorites. The relatively small compositional range of muscovite and paragonite (Table 6) in the metamorphosed samples is also suggestive of an approach towards equilibrium.

Increases in $Al_2O_3$, $K_2O$, Rb, and Zr and decreases in total Fe are apparent between unmetamorphosed and metamorphosed terranes in Tables 3 and 4. Differences in $SiO_2$, $TiO_2$, CaO, MnO, Sr, and Ni, however, are commonly inconsistent between subgraywackes and conglomeratic subgraywackes and commonly show wide variability within a given terrane. The wide compositional variability of the unmetamorphosed rocks indicates that some of the geochemical differences between unmetamorphosed and metamorphosed terranes can be accounted for by original sediment variability. However, anomalously large amounts of sodic plagioclase, quartz, carbonate minerals, and ilmenite in metadiabases, which are intimately associated with the metamorphosed subgraywacke and conglomeratic subgraywacke, suggest that changes in bulk composition also accompanied low-grade metamorphism.

## SUMMARY AND CONCLUSIONS

The late Precambrian rocks of northern Utah and vicinity are composed dominantly of massive, commonly cross-bedded quartzites (70 to 80 percent), lesser amounts of subgraywacke (10 to 30 percent) and conglomeratic subgraywacke (0 to 10 percent), and only minor amounts of carbonate rocks (chiefly dolomite) and metadiabase.

The fact that the subgraywackes are commonly graded, poorly sorted and contain bent and twisted, reworked subgraywacke fragments suggests that most, if not all of them were deposited by turbidity currents. The subgraywackes differ strikingly from glacial varves of the same thickness in that they contain considerably larger amounts of coarse particles and that their carbonaceous matter is not concentrated at the top of each bed. The fact that the conglomeratic subgraywackes contain an abundance of bent and twisted subgraywacke fragments and beds apparently derived from semiconsolidated subgraywacke beds suggests a subaqueous mudflow origin for these rocks. Their intimate association and occasional lateral gradation into poorly sorted, commonly graded subgraywackes suggests that some of the subaqueous mudflows may have given rise to turbidity currents.

Although the textural and structural evidence reported in this investigation suggests a turbidity current and subaqueous mudflow origin for the subgraywackes and conglomeratic subgraywackes, the possibility of contemporaneous continental or alpine glaciation cannot be eliminated because glacial sediments may be redeposited by these agents. Although the large quantity of rounded quartz sand grains in the conglomeratic subgraywackes does not favor a primary glacial origin for these rocks, the possibility still exists of mixing and redepositing nonglacial with glacial material. The large, rounded pebble- to boulder-sized clasts in the conglomeratic subgraywackes may have been shaped and transported to the sites of subaqueous slumping by ice, mudflows, or by streams of greatly varying competency, or by some combination of these agents. The possibility of reworking, mixing and redepositing sediments derived from contemporaneous (and perhaps extinct) glaciers with sediments of nonglacial origin greatly complicates the problem of determining the prevailing climatic conditions at the time these rocks were deposited.

The subgraywackes and conglomeratic subgraywackes are composed chiefly of rounded quartz and minor feldspar sand grains with rock fragments in a diagenetic or metamorphic matrix of mica, chlorite, quartz, and other minor minerals. The original clastic components were probably derived from adjacent older Precambrian terranes in northern Utah and Wyoming. Although the nature and origin of the original matrix minerals are unknown, it is suggested that they were derived from erosion and redeposition of a weathered mantle on plutonic source areas.

Most of the late Precambrian rocks in northern Utah are metamorphosed to the greenschist facies. Large variations in mineral and chemical composition occur within both unmetamorphosed and low-grade metamorphic subgraywacke terranes. Differences in mineralogy and major and trace element concentrations between these terranes appear to be related to a combination of variability in original sediment compositions and compositional changes accompanying progressive metamorphism. Widespread compositional uniformity of low-grade metamorphic plagioclase, chlorite, and, to a lesser degree, muscovite and paragonite, suggests that chemical equilibrium was approached in these phases during metamorphism.

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