PETROLOGY OF THE KEETLEY VOLCANICS IN SUMMIT AND WASATCH COUNTIES, NORTH-CENTRAL UTAH

A THESIS SUBMITTED TO THE FACULTY OF THE GRADUATE SCHOOL OF THE UNIVERSITY OF MINNESOTA

BY

JUSSI EINARI LEVEINEN

IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF MASTER OF SCIENCE

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Abstract

The Keetley Volcanics rest subhorizontally in a structural saddle between the Uinta and Wasatch Mountains ca. 35 km east of Salt Lake City. The Keetley Volcanics consist essentially of Oligocene to Late Eocene volcanic breccias, sandstones, lava flows and porphyritic intrusives. Volcanic breccias are volumetrically the most important, being as thick as 500 m (total) in the vicinity of Keetley. Most of the breccias and intrusives are andesitic. Chemical compositions vary from trachyandesite to silica-poor rhyodacite. The relatively uniform compositions suggest a trachyandesitic source that differentiated by fractional crystallization in a relatively shallow chamber. Large porphyry intrusions are also present west of the Keetley Volcanics. Apparently, the Keetley Volcanics represent a part of magmatic events in the Uinta-Oquirrh mineral belt that started in Late Eocene and continued into earliest Miocene. The Keetley Volcanics may belong to a larger group of mid-Tertiary early to syn-extensional volcanic rocks that were formed before the gradual development of the San Andreas-Basin and Range transform-extensional system.

The primary structures and petrography of the Keetley Volcanics were studied in order to define the mechanics and environments of deposition for volcanic breccias and sandstones that can be formed by several processes. An extensive literature review of volcaniclastic deposits is included.

The studied breccia clasts are almost entirely plagioclase-amphibole porphyritic andesites which commonly contain accessory biotite, clinopyroxene and orthopyroxene phenocrysts. Other accessory minerals are apatite, opaques, sanidine and quartz. Staining for K-bearing minerals indicates that the samples commonly contain sanidine microlites. Other volcanic breccia clasts have a dacitic or basaltic-andesitic mineralogy. Volcanic breccias and sandstones with sand- and gravel-sized dacite clasts indicate a dacitic source for at least some of the distinctively polymictic mass flow deposits. Basaltic-andesitic scoriaceous breccias are associated with basaltic-andesitic lava flows.

The breccias of the Keetley Volcanics commonly include gravel-sized grayish-white clasts that are strongly altered to clay. Microscopic studies indicate that such clasts commonly are pumice. However, some samples have a scoriaceous rather than a
pumiceous texture. Megascopic observation also indicate that some of the clasts are fragments of a pumiceous tuff.

The textures of observed mafic enclaves could not be attributed to magma mixing; together with the common resorbed phenocrysts and rare xenocrysts, they suggest magma/country rock interactions and/or complex crystallization. Slight variations in mineralogy could result from the fact that the breccias were derived from several source rocks or from a single heterogeneous source such as a compositionally zoned magma chamber. Based on mineralogy only, the intrusive rocks cutting the volcanic breccias could have been the sources for the thick breccias.

Because of the narrow mineralogical variation, different andesite lithologies were recognized with difficulty. Porphyritic rocks have been broken into sand-sized clasts that show any combination of one or more phenocryst minerals. Difficulties are also partly due to oxidation, alteration and devitrification. Oxidation has had a stronger influence on the petrography in the basaltic-andesitic clasts than in andesites. Samples of altered or devitrified clasts that have a distinctively different megascopic appearance may appear much less distinctive under a microscope. Different clast lithologies can be defined, based on variations in mineralogy and texture. The results suggest a polymictic and thus, a probable reworked origin for the most of the studied deposits.

The studied outcrops and roadcuts at Jordanelle are located in the Heber quadrangle. The roadcuts are along newly built highway 189. The deposits consist of polymictic to monomictic volcanic breccias and sandstones. Beds commonly have channeled bases and are massive. Inverse or inverse to normal grading is common. Volcanic sandstones and pebbly sandstones are massive or stratified. Some of the stratified lithologies show low-angle cross-bedding. Some of the pebbly sandstones and sandstones are inversely, inversely to normally or normally graded. Volumetrically, sandstones form a minor portion. A few basaltic-andesitic scoriaceous breccia sequences associated with the basaltic andesitic lava flows are present in the upper portion of the sequence which is more than 500 m thick. For many depositional units, a reworked (debris flow or hyperconcentrated flow) origin is probable. However, for some of the depositional units the other alternative, a primary pyroclastic depositional mode, cannot be completely discarded.

The studied outcrops at Indian Hollow are located at the Kamas Quadrangle. The total thickness of the studied sequence is over 100 m. The lower half consists of
andesitic breccias. Thick breccias and conglomerates are laterally persistent but commonly are cut into minor volcanic sandstones and pebbly sandstones, forming gentle channels-fills. Breccia and conglomerate beds are massive or, uncommonly, crudely stratified. Inverse or inverse to normal grading is common. The texture varies from clast-supported to matrix-supported. Most of the breccias and conglomerates are polymictic and contain pieces of columnar-jointed andesites. In a few matrix-supported beds, the clast composition appears monomictic and the sandy matrix is reddish due to oxidation. The beds also contain in situ cooling-jointed blocks. The volcanic sandstones are massive and may show inverse or normal grading. The studied stratified breccia deposits at Indian Hollow are interpreted to have been deposited by lobate- or tongue-shaped debris flows or minor pyroclastic flows. The upper half of the sequence is poorly exposed; it seems to consist of basaltic-andesitic lava flows and scoriaceous breccias.

A sequence of Keetley Volcanics approximately 150 m thick is well exposed at Silver Creek Canyon, 5.5 to 6.5 kilometers southwest of Wanship along Interstate Highway 80 in the Wanship Quadrangle. Polymictic and clearly matrix-supported conglomerates make up an overwhelming part of the sequence. The thicknesses of matrix-supported conglomerates vary from a few meters to several tens of meters. Clast-supported conglomerate beds several meters thick form a minor proportion. The conglomerates commonly contain rounded pebbles and cobbles of older sedimentary rocks. The studied deposits at Silver Creek Canyon can be almost unequivocally attributed to debris flows or hyperconcentrated flows.

Although the Keetley Volcanics rest on the top of volcaniclastic lake deposits (the Peoa Tuff), the studied stratified andesitic to basaltic-andesitic sequences do not support a caldera collapse origin for the Keetley Volcanics. Volcanic breccias and sandstones of the different study areas of this project are similar to debris flow, hyperconcentrated flow and pyroclastic flow deposits that are found at some distance from the source on slopes of modern stratovolcanoes, with the source vent or vents located kilometers to few tens of kilometers from the sites of deposition. The various porphyritic plutons in the Wasatch Mountains 5 to 20 km west of the study area indicate the probable location of the magmatic center and may represent the unroofed remnants of an ancient stratovolcano or an intermediate-silicic multivent center. Some of the material may have been derived from local explosive to gravitational collapses of spines and domes related to shallow intrusions cutting the breccia. Basaltic-andesite
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The time of one year and 8 months I spend in Duluth was a fascinating experience during which I learned a lot, and not only geology. It would be exaggerating to say that it changed the way I think about life but definitely, I got new perspectives. Therefore, I want to dedicate the final word to all of you who made my studies in UMD such an exciting introduction to the American way of life.
Introduction

1. The objectives of the study

Volcaniclastic rocks of the Keetley Volcanics (Kildale, 1951; Bryant, 1990) were studied at five locations (Figure 1) approximately 35 km east of Salt Lake City. The major objectives of this study were to describe the primary structures and petrography of the volcaniclastic rocks and to define the mechanics and the environments of deposition. This work consists of three parts. The first introduces the Keetley Volcanics and their geological setting. The second is a review of volcaniclastic deposits in general and is the basis for the last part which is the description and interpretation of the studied deposits.

2. Geology of The Keetley Volcanics

The Oligocene Keetley Volcanics rest horizontally or subhorizontally in a structural saddle between the Uinta and Wasatch Mountains. The Wasatch Mountains represent part of a Late Cretaceous fold and thrust belt complex uplifted in late Tertiary time during Basin-and-Range block faulting. The Uintas are Paleocene in age and consist of Precambrian rocks that form a broad east-west trending anticline that is restricted by a normal fault along the south flank. This uplift exposed tilted Paleozoic and Mesozoic rocks and it has shed cobbles and pebbles into the unconformably overlying Paleocene-Eocene Wasatch Formation (Bryant, 1990).

The Keetley Volcanics consist of lavas, volcaniclastics and intrusives. The volcanic rocks commonly contain amphibole and plagioclase phenocrysts, as well as less abundant pyroxene and biotite. The Keetley Volcanics have been divided into three lithological units (Bryant, 1990). The lowermost rocks (Tkt) are interbedded fine tuffs, lapilli tuffs, thin lahars and volcaniclastic sandstones and conglomerates. Some of these contain fossils indicating deposition in a lake. These deposits grade upward into a thick sequence of conglomerates and breccias (Tkcb), which form a majority of the volcaniclastic rocks, being as thick as 500 m in the vicinity of Keetley. Lava flows
Quaternary deposits

Keetley intrusives: 1) the Indian Hollow plug; 2) the Park Premier stock; 3) the Jordanelle intrusions; 4) the Provo River intrusion (partly extrusive?)

Keetley volcanics: volcanic breccias, sandstones and lava flows

Tertiary intrusive rocks: granodiorite (5) and granodiorite porphyries (6 & 7) of the Alta stock, monzonite (8) of the Clayton Peak stock

Tertiary rocks - older than the Keetley Volcanics

Paleocene and Precambrian rocks

Figure 1. Geological map of Wasatch, Summit and Morgan counties, north-central Utah. Simplified from Bryant (1990). Study areas (circled) a: Jordanelle; b: Provo river; c: Indian Hollow; d: Francis and e: Silver Creek.
(Tkf) are found within Tkcb; their strike length can exceed 2 km. Age determinations (Bromfield et al., 1977) indicate that The Keetley Volcanics are mainly Oligocene but possibly partly Late Eocene (Table 1). The Keetley Volcanics rest on rocks ranging from the Pennsylvanian Weber Quartzite to the Triassic Thaynes Formation. The wide distribution and variation in thicknesses of basal units suggest rugged relief at the time of volcanic eruptions (Bromfield, 1968).

### Table 1. K-Ar Ages from The Keetley Volcanics (Bromfield et al., 1977).

<table>
<thead>
<tr>
<th>Rock body</th>
<th>Rock type</th>
<th>Location</th>
<th>Mineral</th>
<th>Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flow of Richardson flat</td>
<td>hornblende-bearing rhyodacitic flow*</td>
<td>40°40'52&quot;N 41°26'07&quot;W</td>
<td>hornblende</td>
<td>36.4 ±1.3</td>
</tr>
<tr>
<td>Flow of Richardson flat</td>
<td>hornblende-bearing rhyodacitic flow*</td>
<td>40°40'52&quot;N 41°26'07&quot;W</td>
<td>biotite</td>
<td>33.9 ±1.3</td>
</tr>
<tr>
<td>Park Premier stock</td>
<td>rhyodacite porphyry*</td>
<td>40°38'30&quot;N 41°22'24&quot;W</td>
<td>hornblende</td>
<td>35.2 ±1.0</td>
</tr>
<tr>
<td>Park Premier stock</td>
<td>rhyodacite porphyry*</td>
<td>40°38'30&quot;N 41°22'24&quot;W</td>
<td>biotite</td>
<td>33.9 ±1.2</td>
</tr>
<tr>
<td>Indian Hollow plug</td>
<td>hornblende-bearing latite**</td>
<td>40°38'49&quot;N 41°21'14&quot;W</td>
<td>hornblende</td>
<td>34.2 ±1.3</td>
</tr>
<tr>
<td>Indian Hollow plug</td>
<td>hornblende-bearing latite**</td>
<td>40°39'01&quot;N 41°21'15&quot;W</td>
<td>hornblende</td>
<td>36.1 ±1.3</td>
</tr>
</tbody>
</table>

Rock types from Bromfield et al., 1977 (*) and Woodfill, 1972 (**).
2.1 Intrusive rocks

Intrusion, extrusion and hydrothermal alteration in the Park City area are a part of an overlapping sequence of magmatic events in the Uinta-Oquirrh mineral belt that started in late Eocene, spanned the Oligocene and possibly extended into earliest Miocene (Bromfield et al., 1977). The Keetley Volcanics include several porphyritic intrusions. The largest ones are shown in Figure 1. Large porphyry intrusions are also found west of the Keetley Volcanics in the Wasatch Mountains.

Composition of the intrusive rocks and breccias of the Keetley Volcanics varies from trachyandesite to silica-poor rhyodacite (Bromfield et al., 1977). Generally uniform compositions suggest a trachyandesitic source going through fractional crystallization in a relatively shallow chamber. This trachyandesitic magma was thought to be a result of assimilation of subduction-related calc-alkalic andesitic magmas (Woodfill, 1972). An earlier model by Eardley (1968) involves mantle-derived basaltic source magma. According to Eardley, the Uinta uplift resulted from intrusion of mantle-derived magma into the lower sialic crust. The rising mafic magma caused partial melting and mobilization of the lower crust.

Since these models were proposed, ideas about the tectonics have changed radically. Although the Keetley Volcanics rest between the Uinta and Wasatch Mountains and therefore, slightly east of the Basin and Range region, the Keetley Volcanics possibly belong to a larger group of mid-Tertiary early to synextensional volcanic rocks (Gans et al., 1989, Figure 5).

2.2 Geologic setting

Deposits and magmatism of the western United States record large-scale Cenozoic extension over a 1.500 km wide, zone (Coney and Harms, 1984; Gans et al., 1989; Lister and Davis, 1989). Transform drag of plates started in middle Cenozoic time and has caused the development of a wide tectonic zone. In this zone, faults may have strike slip movement, but they may also indicate regional extension (Basin and Range) or compression (Transverse Range). Nevertheless, there is evidence that extension began 10 to 30 Ma before initial contact between the Pacific and the North America plates and the gradual development of San Andreas-Basin and Range transform-
extensional regimes (Atwater, 1970). Thus, half-graben and metamorphic core complexes were forming in the period between the subduction-related high compression in late Cretaceous and early Tertiary time and the cessation of subduction in the late Tertiary (Coney and Harms, 1984). Several models have been developed to explain the mid-Tertiary tectonic history of western coastal areas of North America.

According to some of the models, Cenozoic tectonism and magmatism were largely directed by plate interactions along the west coast of North America, and early to mid-Tertiary magmatism resulted from the subduction of the Farallon plate. Spatial and temporal variations were controlled by changes in subduction rate affecting the slope of the subducting slab and hence, changing temperature and pressure conditions of partial melting, and the elapsed time of upris ing magma (McKee, 1971; Coney and Reynolds, 1977; Coney, 1979). Models involving extension due to an actively moving oceanic ridge under the continental plate (e.g. McKee, 1971) are, in today's point of view, obsolete since divergence and upwelling at oceanic ridges are not driven by convection currents but instead, by density differences between cooling lithosphere plates and asthenosphere (Hamilton, 1988). Coney and Reynolds (1977) and Coney (1979) favored relaxation of stresses in the overriding plate after a slowdown of subduction rate, allowing extension in the arc or back-arc. It has also been suggested that subduction-related orogeny led to overthickening of crust which then started to extend due to thermal relaxation and gravitational spreading (Coney and Harms, 1984; Wernicke et al., 1987). Relaxation of stresses (Coney and Reynolds, 1977; Coney, 1979) and thermal relaxation combined with gravity spreading are not mutually exclusive and probably both processes played important roles. All these models fail to completely explain some local eruptive and extensional histories or the highly diachronous nature of both volcanism and faulting within the Basin and Range region (Gans et al., 1989).

In the northern Basin and Range province, the onset of both Cenozoic magmatism and extension migrated southward, from Southern Oregon to northern Nevada in the Eocene, to central Nevada and Utah in the Oligocene and to Southern Nevada in the Miocene. On the contrary, in the southern part of the Basin and Range province, migration of magmatism and extension was to the north or the northwest. The onset of dominantly basaltic or bimodal volcanism is also diachronous and precedes the onset of extension by 5 to 15 Ma. Volcanism and extension cannot be explained as a consequence of the migration of the Mendocino triple junction, since the onset of these processes generally predates the migration, especially in the northern Basin and Range.
It seems also that simple Andean-type arc and back-arc extension cannot explain the spatial and temporal variations or the width of the extended crust. Intense extension has taken place in the frontal part of the Laramide thrust belt (southern Nevada) as well as in hinterland areas (west-central Nevada). However, extension was not localized only in previously strongly deformed and thickened areas, so stress relaxation does not seem to be involved. The Laramide orogeny induced prograding amphibolite facies metamorphism in nearly all of the region that was subsequently extended. In many areas there is petrological evidence that the highest temperatures were reached during crustal shortening. If the elevated temperature was required to cause partial melting and to allow extension of thickened crust, spreading should have immediately followed compression everywhere.

In large areas, the onset of extension coincided with or immediately followed extrusions of voluminous intermediate to silicic volcanic magmas (Gans et al., 1989). Despite the fact that the volume of basalts extruded during the early stages is small, several factors suggest that the generation of a large amount of intermediate magma was due to intrusion of mafic mantle-derived magma and subsequent anatexis and magma mixing. For example, the earliest intrusives are commonly the most mafic ones, and are far from minimum granite melts that would be expected from anatexis of overthickened crust. It can be concluded also that the volume of volcanic magma would require unrealistically large degrees of partial melting and would require an external heat source to sufficiently raise crustal temperatures. There seems also to be an increasing amount of petrologic evidence that intermediate mid-Tertiary volcanic rocks were formed by mixing of mantle-derived and anatectic crustal melts. The cause of mantle-derived magma flux remains unresolved. It seems, however, that partial melting of mantle would take place only after extensive stretching of lithosphere. Therefore, the voluminous extrusions of mixed magma at the onset of extension or immediately after it do not support the idea that the flux of mafic magma resulted from extension.

According to Gans et al. (1989), magmatism and associated extension did not solely result from thermal and gravitational instability of overthickened crust, but were triggered and driven by an influx of mantle-derived basaltic magma into the crust. Because the crust was thickened and crustal temperatures were initially close to the solidus of silicic crust, hot mafic magma induced extensive partial melting and consequently, no uncontaminated basalts reached the surface. Anatectic silicic and
mantle-derived magma mixed, hybridized and differentiated, forming compositionally variable small chambers and extrusions. Some of the magma concentrated in larger chambers in which the composition was more uniform (often dacitic) due to thorough mixing. Nevertheless, subsequent fractional crystallization might have caused compositional zoning.

Thermally weakened crust extended and faulting took place in the brittle upper crust. The initial faults were probably localized above the largest accumulations of magma. These initial failures decompressed magma and triggered catastrophic explosive eruptions (Hildreth and Mahood, 1986). Faulting led to the formation of half-graben complexes which, as suggested by thick sequences of mid-Tertiary fanglomerates and lacustrine deposits, had geometries and dimensions comparable to those of the present Basin and Range province.

Localized extensional strains were accommodated by broader stretching and flow at lower crustal levels. Strain rates decreased exponentially with gravitational potential of thinning and stretching crust until the present, relatively stable crustal thicknesses were reached. Late extrusions were relatively uncontaminated because in the extended crust the proportion of cold, brittle crust was greater (Gans et al., 1989).

Mid-Tertiary magmatism occurred also in areas which obviously were not significantly extended. This suggests that the influx of mantle-derived magma and associated volcanism was widespread but extension was triggered only in relatively narrow areas that were previously weakened. Models for extensional tectonism are in a developing state and very likely the model of Gans et al. (1989) will not represent the final opinion.
Volcaniclastic sedimentology: literature review

3. Terminological considerations

The term "volcaniclastic" includes all clastic materials containing significant proportions of volcanic fragments formed by any fragmentation process. Volcaniclastic particles (i.e., volcaniclasts) can be formed by several processes. Pyroclasts form from expansion of gases within a magma that loses its cohesion and breaks into bits when overburden pressures are exceeded. Hydroclasts are formed by magma-water interactions that are either explosive or nonexplosive. Epiclasts are particles derived from any type of pre-existing rock by weathering and erosion. Autoclasts are formed by mechanical breakage of lava flows due to friction or by gravity crumbling of spines and domes. Alloclasts are fragments of pre-existing volcanic rocks broken by igneous processes beneath the earth's surface. Deposition of volcaniclastic material can result directly from eruptions as fall-out, flows and surges. Volcaniclastic material can also be carried to sites of deposition by wind, rivers and various mass transportation processes. During transportation, volcaniclastic material can have one or more depositional stages interrupted by erosion and retransportation. Reworking and recycling of un lithified and lithified volcaniclastic material results in a wide spectrum of deposits.

Lajoie (1984) described pyroclastic fragments as primary if ejected from a vent to the site of deposition, but secondary if the pyroclastic material was redeposited before consolidation.

According to the classification of Cas and Wright (1987), pyroclastic deposits refer to aggregates that are not only fragmented by explosive activity but also deposited by transport processes resulting directly from this activity. They also used the term epiclastic to describe deposits that were "produced by normal surface fragmentation processes (weathering, physical abrasion, gravitational collapse) or were finally deposited by normal surface processes (traction, suspension, mass flow) irrespective of their fragmentation mode, or both" (Cas and Wright, 1987, p. 8). Thus, in this classification the transportation mode of deposits is emphasized. Volcaniclasts are
produced in large volumes and instantaneously generate facies distributions and geometries not met in non-volcanic epiclastic deposits. Volcanism and volcaniclastic deposition are contemporaneous and cogenetic processes and, according to Fisher and Smith (1991), it is important to separate clearly the fragmentation processes from the transportation processes. In order to do this, Fisher and Smith (1991) defended their classification, according to which epiclastic deposits are made of weathered and eroded fragments from pre-existing rocks, and pyroclastic deposits consist of pyroclasts whether of primary or reworked origin. Both of these approaches have been used. The unfortunate result of this is that such a fundamental term as "pyroclastic deposit" can have different meanings (Fisher and Schminke, 1984; Cas and Wright, 1987; Fisher and Smith, 1991).

It is especially difficult in ancient deposits to determine the fragmentation processes or the depositional mode. Therefore, even without the above-mentioned controversy, a descriptive non-genetic classification is needed. The non-genetic classification of volcaniclastic deposits used in this paper is the one suggested by Cas and Wright (1987). Beds are described following Ingram's (1954) classification as laminae (< 1 cm) and very thin (1-3 cm), thin (3-10 cm), medium-thick (10-30 cm), thick (30-100 cm) or very thick beds (> 100 cm)

Pyroclasts and pyroclastic deposits are commonly classified according to Table 2. In this classification a block is "a pyroclast, more than 64 mm in diameter that was ejected in a solid state" (Bates et al., 1984) and that is therefore angular. In sedimentological literature, a block is commonly an angular boulder-sized clast (Ø > 254 mm) "showing little or no modification by transporting agents" (Bates et al., 1984). In most cases the fragmentation process cannot be defined. Some deposits contain both epiclasts and pyroclasts. In this study of the dominantly coarse-grained rocks such as the Keetley Volcanics, a nongenetic grain-size classification is needed. Then, if a pyroclastic origin in the sense of Cas and Wright (1987) or Maitre (1989) can be established, the classification and nomenclature of Maitre (1989, p 7-9) has been used. To describe all the volcaniclasts, grain sizes are classified according to the Udden-Wentworth scale.
Table 2. Classification of pyroclasts and pyroclastic deposits based on clast size (Schmid, 1981; Maitre, 1989).

<table>
<thead>
<tr>
<th>Clast size (mm)</th>
<th>Mainly unconsolidated:</th>
<th>Mainly consolidated:</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>tephra</td>
<td>pyroclastic rock</td>
</tr>
<tr>
<td>bomb/block</td>
<td>agglomerate/</td>
<td>agglomerate/</td>
</tr>
<tr>
<td>64</td>
<td>block tephra</td>
<td>pyroclastic breccias</td>
</tr>
<tr>
<td>lapillus</td>
<td>lapilli tephra</td>
<td>lapilli tuff</td>
</tr>
<tr>
<td>coarse ash grain</td>
<td>coarse ash</td>
<td>tuff</td>
</tr>
<tr>
<td>1/16</td>
<td>fine ash grain</td>
<td>fine tuff</td>
</tr>
</tbody>
</table>

4. Theoretical considerations

4.1 Pyroclastic flows and surges

Eruptions generate down-slope moving mixtures of juvenile pyroclasts, magmatic volatiles and external water. These mixtures have traditionally been divided into high concentration pyroclastic flows and dilute surges. Pyroclastic flows and surges move under the influence of gravity with speeds varying from tens to hundreds of meters per second. It is somewhat disputable whether or not there is a complete transition (Cas and Wright, 1987) or distinctive gaps (Walker, 1983) between these two end members. Although differences are clear at both ends of a spectrum, there are many pyroclastic flow deposits that cannot be conveniently placed into either category. Pyroclastic flows and surges are complex depositional events in which depositional mechanisms are temporally and spatially variable. During each event, a set of transportation and depositional processes produces a wide diversity of deposits (Carey, 1991).
4.2 Pyroclastic surges

Pyroclastic surges are low concentration mixtures of volcanic fragments, gases and, to some extent, water, that move in turbulent fashion as gravity flows. Three different types of surges has been recognized. Base surges form radially spreading clouds from the base of a phreatomagmatic eruption column. Ground surge beds are found at the bases of pyroclastic flows. Ash cloud surges are found covering pyroclastic flows. The origin of ground surge deposits is uncertain although they are undoubtedly related to pyroclastic flows. According to Sparks and Walker (1973), a dilute cloud is segregated from an initially homogenous pyroclastic flow. This cloud can eventually leave behind the slowly moving basal part which is then overridden by the flow. Fisher (1979) suggested that ground surges result from an initial collapse of the fine-grained outer parts of eruption column. According to other theories, ground surges are deposited from overpressured blasts (Wohlets et al., 1984) or from turbulent heads and bases of pyroclastic flows (Valentine and Wohlets, 1989; Valentine and Fisher, 1986). Ash cloud surges are deposited from dilute clouds of elutriated ash that move slower than the main flow (Cas and Wright, 1987; Carey, 1991).

4.2.1 Pyroclastic surge deposits

Pyroclastic surges are formed from dilute flows that are not strongly restricted by topography. They tend to form more radially distributed deposits around the vent. Individual deposits are thicker in valleys and indentations, but can mantle topographic heights and obstacles. Carbonized wood is associated with some surge deposits. Some have accretionary lapilli or are plastered around obstacles, indicating the presence of water and low temperatures. Characteristic features of some surge deposits are low-angle dune-like bedforms and cross-bedding. Surges can also contain planar stratification or massive beds which, especially when thin and distal, can resemble pyroclastic fall deposits (Sigurdson et al., 1987; Cas and Wright, 1987; Carey, 1991). In ground surges, dune-like bedforms are found at proximal distances (several kilometers) changing to massive and finally into distal stratified deposits (tens of kilometers). In stratified deposits, individual layers are well-sorted in terms of fall velocity, even if sorting is poor in the overall sequence. In the plots of sorting versus median diameter, surge deposits represent an intermediate field between fall and flow.
deposits but because of the large overlap the grain-size characteristics of surge deposits are not distinguishing features.

The shapes of the dune-like bedforms are strongly influenced by flow velocity and moisture content (Allen, 1984). Crest lines tend to be perpendicular to flow direction. Regressive bed forms migrate toward the source. The migration takes place by the growth of stoss-side laminae. Lee side laminae are thinner and have gentler dips. They are preserved if the deposition rate is high compared to the transportation rate. The formation of regressive bedforms is favored by relatively high moisture contents and low temperatures.

Progressive bedforms migrate away from the source by depositing steeper and thicker laminae on the lee side of the bedform. Stoss-side laminae are only preserved if the sediment deposition rate is high compared to the transportation rate. The formation of bedforms is favored by relatively dry and/or hot conditions. Stationary bedforms have symmetrical cross-sections and similar dips on both sides of the crest. They represent intermediate flow conditions (Allen, 1984).

4.2.2 Depositional processes in pyroclastic surges

Analogy with other types of mass flows suggests that in surges the particles are strongly supported by turbulence and that the clast concentration is relatively low. Turbulence is maintained by transformation of gravitational potential energy to kinetic energy while the internal friction is small due to the low concentration of particles. Fluidization is probably not important because of the high gas velocities that are needed to completely fluidize surge material. Such high velocities would lead to the extensive removal of fine ash and very good sorting. Instead, better sorting compared to pyroclastic flows can be attributed to lower concentration and high turbulence (Sigurdson et al., 1987).

Surges show similarities to turbidity flows moving downslope as high density suspensions through low density fluids (Allen, 1984). The main difference is that surges move through air and that the fluid can be composed of multiple phases of gas and water. These gases can expand supersonically, contributing to the high kinetic energy of surges. As the flow moves, it responds to gravity by segregating larger particles towards the base. Segregation leads into layered flow with material exchange between
the layers. The main part of the surge is turbulent and dilute. In the basal zone, clasts are transported by traction, saltation, rolling and sliding. Basal transportation also involves thin layers or traction carpets of thin grain flows moving in a pulsative fashion (Carey, 1991).

Dune-like bedforms of surges have been compared to fluvial dunes. Nevertheless, especially in the case of wet surges, cohesion may have an even greater influence on bedforms than flow velocity (Allen, 1984; Carey, 1991). Bedforms of hot and dry surges are probably comparable to eolian dunes. Avalanching causes inverse grading in lee-side laminae of eolian dunes. Similar structures have also been found in surge deposits (Allen, 1984; Sigurdson et al., 1987). Surges, however, have higher concentrations, flow velocities and variance in grain sizes. These factors induce traction carpets of pulsating grain flows. Movement of such flows produces low-angle cross-bedding that is typically found in surges (Cas and Wright, 1987). Valentine (1987) has suggested that the decrease in wave length with distance from the source can be explained by a decrease of maximum wave length of internal wave propagation in stratified flow.

5. Pyroclastic flows

Pyroclastic flows are generated by eruption column collapses, explosive or gravitational collapses of a summit dome or overspilling out of a crater (Cas and Wright, 1987; Smith and Roobol, 1990; Carey, 1991). The resultant deposits can be divided into different categories based on vesicularity and the composition of juvenile clasts (Cas and Wright, 1987). Pumice flows or ignimbrites have abundant low density pumice, usually with a high silica composition. The pumice flow matrix contains large quantities of fine ash. If fine ash is the dominant component (over 50 percent), flows can be called ash flows.

Scoria flows contain abundant scoriaceous material having a basaltic to andesitic composition. Some dense cognate lithic clasts are usually present.

Block and ash flows are characterized by large dense or sparsely vesicular blocks. The term "block and ash flow" can be misleading but remains widely used. In finer flows, block-sized fragments are not necessarily present and thick and coarse deposits
can be strongly fines-depleted, containing little ash. The name can also give an impression of bimodal grain-size distribution, which is not always the case (Smith and Roobol, 1990).

Volumes of individual flows range from fractions of a km$^3$ to several thousands of km$^3$. Pyroclastic flows move under the influence of gravity and therefore their morphology and distribution is effected by topography. The scale of topographic effects depends on the volume and momentum of a flow. Small- and intermediate-sized flows can be restricted into river valleys, and ridges and barriers can halt or divert their progress. Large flows tend to form sheet-like bodies and can completely overwhelm previous topography.

Existing observations provide little detailed information about particle transportation, flow regime and flow structure. These characteristics are studied by comparing pyroclastic flow sequences to deposits of other particulate gravity flows. The studies of vertical and lateral variations of pyroclastic flows have led to several standard sequences, as described below.

5.1 Pumice flows

5.1.1 High aspect ratio pumice flows

The sequence for ignimbrites (Sparks, 1976) consists (in the ideal case) of three distinct units. The lowermost unit (layer 1) consists of stratified and commonly cross-bedded surge deposits. The main unit (layer 2) is a thick pyroclastic flow unit that is commonly covered by thin ash deposits (layer 3). The main unit is very poorly sorted, containing meter-sized clasts as well as glass shards only a few microns in diameter. The unit lacks internal stratification but shows systematic gradual changes in grain-size distribution. Near the base, both pumice and dense lithics are inversely graded. This inversely graded subunit (layer 2a) is also found along steep valley walls. The basal subunit grades upwards into a subunit in which low density pumice and high density lithics show the opposite sense of grading. Grading patterns are also a function of a grain size. The bigger the blocks, the more susceptible they are to concentration towards the top (pumice) or near the base (lithics) of layer 2b. With decreasing grain size, the tendency to grade decreases and finally, the proportion of fine ash (< 0.5 mm)
is uniform irrespective of the vertical position. Therefore layer 2b shows coarse-tail grading that is inverse for pumice but normal for dense lithic clasts.

Not all pyroclastic flows show similar grading patterns. Many flows, even pumice flows, are ungraded (Cas and Wright, 1987) or show normal grading of both pumice and lithics (Carey, 1991). The grading patterns of pyroclastic flows are strongly related to initial grain-size distribution and density contrasts of different components. More or less ideal sequences are derived from medial to distal parts of deposits that have generally high aspect ratios (thickness/length $10^{-2}$ to $10^{-3}$).

5.2.1 Low aspect ratio pumice flows

Low aspect ratio (thickness/length $10^{-4}$ to $10^{-5}$) pyroclastic flows seem to be more complex. An example of a low aspect ratio pumice that records deviating facies relationships is the Taupo Ignimbrite from New Zealand (Walker et al., 1981; Wilson and Walker, 1982). Layer 1 consist of two clearly different varieties which are separated from overlying layer 2 by a sharp boundary. The other is almost entirely pumice-rich and the other is lithic-rich. The pumice-rich layer (1P) may be overlain by the lithic rich variety (1H). The pumice-rich layer is somewhat depleted of fines compared to layer 2 deposits. Layer 2 also seems to have two varieties. Valley-pond ignimbrites (VPIs) are commonly from 1 m to 70 m thick and consist of one depositional unit, often having similar grading patterns to layer 2 of Sparks (1976). In some cases, the concentration of pumice towards the top has led to relatively sharply bounded pumice concentration zones which contain the largest pumice and practically lack other components.

Deposits mantling ridges in between VPIs are characteristically layered due to variations in contents of coarse pumice and slight changes in grain-size distribution. Yet, the general grain-size distribution in these ignimbrite veneer deposits (IVDs) is similar to those in VPIs, except that the maximum grain size is smaller. The major difference between VPI and IVD deposits is that the former are 5 to 50 times thicker than the latter. Field observations confirm that VPIs and IVDs are laterally equivalent deposits of single volcanic events. Facies relationships are not only dependent on topography but are also dependent upon distance. The facies sequences described above are found at moderate or great distances from the source vent; proximal parts of pyroclastic sequences seem to be characterized by various types of breccias.
5.2 Scoria flows and block and ash flows

In scoria flows, large blocks are typically inversely graded and gas escape structures may be present. Flows may be preceded by fine stratified deposits (layer 1) and fine-grained basal layers may have developed (Cas and Wright, 1987).

Thick breccias composed of dense or sparsely vesicular blocks are prominent deposits associated with classic Peléean type eruptions named after Mt. Pelée, Martinique (Cas and Wright, 1987; Smith and Roobol, 1990). Peléean type volcanism involves an actively growing dome and the generation of block and ash flows. Recorded historic events have taken place with five different mechanisms. A growing dome may collapse due to gravitational instability. Collapse may be also a result of increasing gas pressure causing an explosion of a dome. Examples from other volcanoes suggest that initial gravitational collapse of a dome can expose gas-rich magma in the interior of a dome. As the gas explosively expands, the difference between gravitational and explosive collapse disappears. Flows also result from the vertical explosion of a dome followed by collapse of the eruption column. Observed lateral explosions from near the base of a dome are probably similar to the May 18, 1980 blast of Mt. St. Helens, but smaller in scale. Some flows are probably extrusions from around the base of a growing dome that spilled over the crater rim without a significant eruption column (Smith and Roobol, 1990).

Block and ash deposits in the main fan of Mt. Pelée are typically angular to subangular. Diameters of blocks can be as much as 6 m although the average diameter is much smaller. Larger clasts (> 60 cm) commonly have very often well-developed postdepositional radial cooling-joints or, occasionally bread crust structure, indicating emplacement by a hot flow and cooling in situ. Blocks are essentially derived from one source. Coarser clasts are dense to sparsely vesicular juvenile material, derived from growing dome. Altered fragments of older rocks are sparse. Vesicular juvenile clasts are seldom larger than 30 cm in diameter. Sparse lithic clasts are usually less than 3 cm in diameter. Distribution of clast components reflects mechanical controls. Accidental lithic fragments were gathered by viscous vesiculating magma from vent walls and carried through a dome. Larger fragments were formed during dome collapse.

Block and ash flow deposits commonly show only crude bedding, resulting from variations in sizes of the larger clasts. Layering is often indicated by fine-grained basal
layers and interbedded surge deposits. Individual flow units are 1 to 20 m thick in the Mt. Pelee main fan. Flow units almost invariably have fine-grained bases, comparable to layer 2a of Sparks (1976). Fine-grained layers similar to layer 1 (Sparks, 1976) are missing. The main part of a block and ash flow can be inversely or normally graded. Some deposits show complex channeling and pockets of smaller blocks on the vent sides of blocks. Complexities may represent separate flow units or flow pulses. Segregation lenses of blocks may be attributed to internal variations in shear stress distribution.

Gas segregation pipes are common in the deposits of Mt. Pelee (Smith and Roobol, 1990), but these are not commonly found in block and ash flows (Cas and Wright, 1987). Escape structures penetrate both block and ash deposits and the surge deposits that may follow and which are probably deposited rapidly from ash clouds, subsequent to block and ash flows.

Near-vent block-dominated block and ash flow deposits of Mt. Pelee grade into and intercalate with lag breccias (described below) of juvenile blocks. In distal parts, the proportion of blocks decreases as the competence of flows to transport coarser material diminishes. Approximately 8 km from the vent of Mt. Pelee, ash-size material predominates (Smith and Roobol, 1990). Blocks form only occasional lenses in which coarser fragments are clast-supported. In addition to longitudinal changes, the deposits change laterally to thinner and finer (maximum Ø 70 cm) deposits outside the main fan. Changes can take place within a few meters. Finer and thinner lateral equivalents have been interpreted in two different ways. One explanation favors detachment of a cloud containing some coarser material from the main flow and subsequent resegregation of the ash cloud into a coarser basal layer and a finer ash-rich cloud. Such deposits could also represent pyroclastic flows that spilled over confining valley walls and spread out laterally.

5.3 Transportation mechanisms of pyroclastic flows

The mobility of pyroclastic flows is a function of eruption intensity and volume. The major controlling factor seems to be the height of the collapsing eruption column. Although the elevation of the vent contributes to the potential energy of a flow, high mobility flows can be formed from vents that are on flat areas such as at Lake Taupo,
New Zealand (Carey, 1991). Among pyroclastic flows, pumice- and ash-rich flows are more mobile than flows dominated by dense lithic fragments. The former type of pyroclastic flows have been able to surmount barriers up to 600 m high and they clearly have a higher mobility than other types of granular mass flows. Small volume flows, composed mostly of dense lithic fragments, are no more mobile than cold avalanches.

The ability of pyroclastic flows to override high eminences was originally thought to indicate that pyroclastic flows were highly expanded mixtures of debris and gas. Pyroclastic flows are usually not extensively thicker than the deposits they leave behind. The controlling factor seems to be the momentum of the flow.

The major clast-support mechanism during the flow is the fluidization by expanding and escaping juvenile gases. Heating and expansion of entrained air may also increase the mobility of pyroclastic flows. Fluidization is a process in which debris loses its strength due to gases percolating through intergranular pores. If the grain size and/or density range is narrow, a gas flux causes slight expansion until a critical gas velocity (minimum fluidization velocity, $U_{mf}$) is reached. Then the drag force exerted throughout the bed is equal to the buoyant weight of the bed. The bed loses its strength and adopts a fluid-like flow behavior. If the grain size range or density range or both are wide, as the case is with pyroclastic flows, the bed can be only semi-fluidized (Sparks, 1976). When gas velocity ($U$) becomes enough high to fluidize larger or denser clasts, it has already exceeded the terminal fall velocity of finer grains ($U_t$). These finer grains are thus removed from the debris by escaping gas. This process is called elutriation. According to Sparks (1976), three phases are present in a pyroclastic flow. The first phase comprises the fine grains of the flow matrix which have $U_t$ values smaller than the sustaining gas velocity $U$. These grains are elutriated from the flow matrix. The second phase involves fluidized grains which have $U_{mf}$ less than or equal to $U$ but which are not removed by elutriation ($U$ smaller than $U_t$). The third phase is composed of dispersed clasts that are not fluidized ($U_{mf} > U$). These clasts either sink or float, depending on their density compared to the density of the fluidized matrix (Sparks, 1976; Cas and Wright, 1987).

Fluidization explains observed grading patterns, the high mobility, and the ability to carry large clasts (layer 2b, Sparks, 1976). This process requires that the density of matrix is between that of the pumice and lithic clasts. Therefore, even when moving
fast, fluidized pyroclastic flows are not largely expanded compared to the thickness of resulting deposits. Thus their density is high enough to render possible the support of large clasts.

Near the base of a pyroclastic flow, clasts are also supported by dispersive stress (discussed later), acting perpendicular to the lower bounding surface. This stress is induced by basal shear and it strongly supports larger clasts, causing inverse grading within the basal layer (2a, Sparks, 1976).

Sparks' (1976) results were based on equations developed for purposes of chemical engineering. The process of fluidization seems to be more complex in real pyroclastic flows and involves the formation of gas segregation pipes when gas flow exceeds a threshold value (Wilson, 1980, 1984). According to Wilson (1980) three different fluidization degrees can be recognized. Flows having the lowest fluidization stage lack grading of large clasts. Their matrix contains fine particles indicating ineffective elutriation and expansion of flow matrix. The second fluidization stage flows show coarse-tail grading that is inverse for pumice and normal for lithics. In these flows, gas flux is adequate for the expansion of the flow and partial fluidization of the matrix. Larger clasts are able to either float or sink, depending upon the buoyancy force. Flows representing a high stage of fluidization are depleted in fine ash fraction due to elutriation. They are also characterized by gas segregation pipes and pumice-rich layers at the top and lithic-rich ones at the base and of the flow due to bubbling of the gas-debris mixture (Wilson, 1980, 1984).

5.4 Deposition of pyroclastic flows

Pyroclastic flows show several similarities with various high density sedimentary gravity flows (Sparks, 1976; Wilson and Walker, 1982) such as poor sorting, variable grading patterns, matrix-support of large clasts and the presence of fine-grained basal layers. Internal structures of flow deposits reflect clast-supportive forces (e.g. fluidization, dispersive stress) that were active during transportation. Similar to sedimentary debris flows, the depositional mechanism in many pyroclastic flows is probably rapid in situ freezing. Freezing takes place when the yield strength exceeds induced shear. The yield strength can be attributed to cohesive and frictional forces.
between the clasts (Sparks, 1976). Resulting deposits are frozen remnants of the moving flow, and their grain-size distribution represents that of the flow itself.

Many flows having low aspect ratios and more expanded and dilute flow conditions show similarities to more dilute sedimentary gravity flows such as turbidites. Different depositional modes seem to have occurred during the deposition of single flow units. More complex facies relationships result from segregation of the flow into head, body and tail (Wilson and Walker, 1982; Carey, 1991).

In the head of a pyroclastic flow, gas is added by entrapment of air and by combustion of buried vegetation leading to high fluidization. High gas velocities and fierce turbulence cause the settling of dense clasts towards the base and elutriation of fine ash. Deposition of segregated material forms the vitric clast-rich layer of low aspect ratio flows (IH layer, Walker et al., 1981) that is overridden by the rest of the flow. Eyewitnessed flows have had jets of heated and expanded trapped air puffing out of the head. The stratigraphic position and fine ash-depleted grain-size distribution suggests deposition of the pumice-rich layer of low aspect ratio flows (IP layer, Walker et al., 1981) from such turbulent and low concentration jets. Deposition from main parts of highly expanded, low aspect ratio flows, such as the Taupo Ignimbrite, are still under debate. The ignimbrite veneer deposits (IVD) described by Walker et al. (1981) have features of both low and high aspect ratio pyroclastic flow types. They are stratified-like surges but are poorly sorted like pyroclastic flows. Deposition takes place layer-by-layer, mainly from the tail of the flow. The layering of IVD deposits reflects variations in flow thickness at a site of deposition that caused changes in internal shear and deposition rate. Variations of flow thickness result from discrete flow pulses or waves on the upper surface of the flow. The grain-size distribution is finer in IVDs (Fisher, 1979; Wilson, 1985; Carey, 1991).

Valley pond ignimbrites have internal structures typical of moderately expanded flows (poor sorting, lack of stratification, pumice concentration zones). On the other hand, there is evidence that the flows that deposited VPIs were much thicker than the resultant deposits. This paradox is explained by Wilson and Walker (1982) to be a combination of thicker flow tails in valley floors, drainage of the upper parts of the flows into the valleys, and blockage by logs and large pumice at valley constrictions.
Facies relationships of pyroclastic flows seem to be a function of eruption intensity. High-eruption energy flows move faster; the entrapment and heating of air significantly increases fluidization. The lowermost deposits result from segregation and jetting at the flow front. Facies relationships are derived from fractionation of a flow into head, body and tail. Deposits grade laterally to finer and thinner deposits. Deposition ceases when the amount of debris diminishes. At low velocities, entrapment of air is insignificant and segregation and jetting deposits (1H and 1P, Walker et al., 1981; Wilson and Walker, 1982) are not formed although there might be surge deposits (discussed later) underlying the main flow, forming an association of layers 1 and 2 (Sparks, 1976). Deposits represent frozen remnants of the flow. Movement stops due to yield strength, and deposits have blunt fronts and edges.

5.5 Pyroclastic breccias

Pyroclastic breccias are characteristic near vent areas. Breccias can be formed by several processes. Unfortunately, distinguishing them from other types of volcanic breccias resulting from debris flows and other types of laharc-related mass flows, debris-avalanches and rock fall deposits, can be very difficult.

Pyroclastic breccias can be coarse fall-deposits. Thick breccias are formed by caldera collapses. Proximal block and ash flows are examples of pyroclastic breccia. Breccias are also commonly associated with pumice flows. Therefore, such deposits are called co-ignimbrite breccias. Two different types of flow-generated co-ignimbrite breccias can be recognized. One type is followed by a main body of a pyroclastic flow showing 2a and 2b-layers. These deposits are a coarse variation of a ground layer (layer 1, Sparks, 1976). The breccia layer is separated by a sharp and near planar sheared contact from the layer 2 ignimbrite. Deposits are commonly less than 1 m thick. According to Walker (1985), a complementary relationship exists between the ground layer and layer 2 of the ignimbrite. In the ground layer the mean size is 3 or 4 times that of the layer 2 ignimbrite at the same outcrop. Ground layers are also typically strongly fines-depleted.

Another type of breccia is a lithic concentration zone at the base of layer 2b. Such breccias are seldom thicker than 1 m. Breccias are more commonly matrix-supported than clast-supported. Beds may be graded, but internal stratification is lacking. The
matrix contains some fines. The lithic clasts may have been segregated from the pyroclastic flow or remobilized by the flow from underlying deposits. Therefore the clast-composition may be different from the main part of the flow (Walker, 1985).

Both ground layer and layer 2b breccias are formed in pyroclastic flows in which segregation of the gas phase is in progress. Segregation becomes significant at some distance from the vent. Close to the vent, other types of co-ignimbrite breccias are formed. These are called lag-fall breccias.

Lag-fall breccias (Walker and Wright, 1977; Walker, 1985) are characteristically fines-depleted, clast-supported and stratified. Breccias have been "deposited by the addition of layer upon layer from the bottom up as by a kind of fall mechanism" (Walker, 1985, p. 157). Some characteristics however, suggest a simultaneous lateral flow component. Such beds are commonly discontinuous and tend to thicken into valleys. Some beds have erosional bases (Walker, 1985). Beds are typically 1-20 m thick and are coarse-grained with clasts predominantly coarser than 16 mm. The presence of large blocks (Ø > 50 cm) indicates a high flow competence. Although characteristically fines-depleted, the breccias may contain smaller pumice lapilli. Apparently during deposition, while blocks are too heavy to be transported further, fine ash can be removed by elutriation. Some lapilli are trapped and lodged between larger blocks. Indistinct stratification of the deposits suggests deposition from a vigorous eruption column with slight variations in eruption intensity.

Lag-fall breccias are lateral equivalents of coarse fall deposits near vent areas and down slope from pyroclastic flows. Deposits described by Walker and Wright (1977) show compositional zoning similar to synchronous ignimbrite flows. Near-vent ("pure") fall breccias contain abundant bombs and ballistic impact structures. Deposits are better sorted than lag-breccias and lack smaller clasts. Ground layers of proximal pyroclastic flows are similarly fines-depleted but layers are typically thinner, finer-grained, and lack internal stratification. This fines-depletion distinguishes lag-fall breccias from caldera collapse breccias. Lag-fall breccias are commonly formed at the eruption climax, particularly at the onset of caldera collapse. Lag-fall breccias can be deposited at distances of 5-20 km from vents and can occur as interbeds in thick pumice flow sequences (Walker, 1985).
6. Deposition of gravelly deposits

6.1 Deposition by free-fall

Particle free-fall may occur subaerially or subaqueously near steep slopes or subaqueously by drifting vegetation carrying rocks in root systems or by drifting ice. Repeated deposition onto a hard substratum leads to an unstructured pile of angular fragments. Fall may occur into soft sediment, causing deformation and impact structures. Rock piles deposited on breaks of steep slopes are laterally discontinuous and closely associated with their source walls. Drop-stone deposits from drifting trees and ice typically have heterogeneous clast compositions, missing easily abraded material, such as pumice or scoria, and they have variable clast shapes. Their matrix content and facies association usually helps to differentiate them from pyroclastic fall deposits containing ballistic ejecta. Rock fall may reach the mass movement mode if a large volume of rock falls grades laterally into air-lubricated debris avalanches or viscous debris flows (Jones et al., 1984).

6.2 Deposition by avalanches

6.2.1 Triggering of avalanches

Avalanches are the result of large-scale failures of volcanic cones or slopes, leaving horseshoe-shaped or amphitheater-like scars. Such huge landslides may have displaced half of a shield volcano in Hawaii (Holcomb, 1983) and their volumes may have exceeded 250 km$^3$ (Sundel, 1983). Usually the volume of deposits exceeds 1 km$^3$. Avalanches are very mobile. Calculated emplacement velocities usually exceed 100 km/h and deposits can be traced for distances of 50-60 km.

Failures are usually triggered by earthquakes or explosive eruptions. A directed blast may be also be a result of an avalanche, as was the case during the May 18, 1980 eruption of Mt. St. Helens (Voight, 1981). The instability of the volcanic edifice may be promoted by 1) dilation around shallow intrusives or parallel dike swarms 2) massive
lavas overlying a pyroclastic substratum 3) hydrothermal alteration weakening a volcanic cone 4) the migration of an eruption vent or 5) the reduction of strength due to saturation by water (Siebert, 1984).

6.2.2 Characteristics of avalanche deposits

Avalanche deposits commonly have well-formed levees. Their most characteristic morphological features are numerous hills and transverse and longitudinal ridges forming a hummocky topography. There can be thousands of hills associated with a single avalanche. The hills can be several tens of meters high and their diameters can be several times their height. Hills are often unequally distributed so that clusters of hills may be surrounded by relatively flat areas. Hummocks (hills) are probably results of horst and graben formation during lateral spreading (Siebert, 1984).

Inter-hummocky areas may be also filled with secondary debris flows and/or other reworked materials. The maximum size of hummocks tends to decrease towards the distal part of a flow, but in many deposits the largest hummocks are found in the central portions. A slope failure may take place as subsequent smaller failures and additional debris filling, causing an apparent decrease in proximal hummock size (Ui, 1983; Siebert, 1984).

Debris avalanches are massive and disorganized. They are coarser grained and more poorly sorted than pyroclastic flow deposits. Avalanches have been described with very variable terms. Avalanches are capable of carrying huge blocks of coherent, unconsolidated or poorly consolidated pieces of a volcano. Such blocks may have diameters as large as 500 m. Such blocks are called megablocks (Ui, 1983). Since they can be any size, the term "debris-avalanche block" was adapted by Glicken (1991). Hard pieces of rock carried by avalanches are referred to as "clasts". Clasts and debris-avalanche blocks may be surrounded by finer matrix (< 2 mm). Interclast matrix is between clasts or debris-avalanche blocks; intraclast matrix is found within blocks (Palmer et al., 1991). Debris-avalanche blocks may preserve their original layering or bedding. Paleomagnetic measurements indicate that blocks are rotated only horizontally during transportation. Adjacent debris-avalanche blocks can often be fitted together in a jigsaw-puzzle manner. Clasts are characteristically fragmented in situ. Fragmentation can result in the complete shattering of all the clasts.
Debris flow deposits may carry exotic blocks but their matrix is homogeneous and thoroughly mixed. "In debris avalanche deposits large blocks are surrounded by matrix of similar composition, which may be in sharp contact with blocks and matrix of a different composition" (Siebert, 1984, p. 167).

Poor mixing of matrix and the preservation of magnetic orientation suggest laminar plug flow instead of turbulent flow. The angular and poorly sorted texture of matrix, sharp changes in its composition, and down-slope decrease in maximum grain size, suggest that the matrix is largely formed by crushing during transportation (Ui, 1983).

The friction at the bottom of avalanche deposits is extremely low and it has been suggested that avalanches leave the ground. The contribution of trapped air to mobility and clast-support is not completely solved. Apparent low friction might be caused by the overburden compressing a cushion of trapped air. However, since there are similar deposits on the moon, the interstitial fluid or compressed gas is not necessary for motion. The support is therefore dominantly driven by gravitation and is due to collisions of blocks in fine debris and dust causing dispersive stress (Bagnold, 1954; Howard, 1973; Ui, 1983). As opposed to grain flows (Lowe, 1982) which are driven by dispersive stress as well, the friction between the flow and the substratum is apparently low. The reason for this remains unknown.

Segregation or acquisition of water can change flow characteristics of avalanches so that they can be gradational into debris flows of volcaniclastic material. This can happen if hot interstitial gas condenses, if an avalanche reaches a water body, or if ice and snow carried by the avalanche melt due to impacts. Features of avalanche deposits may be gradational with those of pyroclastic flows, reflecting the continuum between non-explosive failure of slopes to explosive magmatic or phreatomagmatic eruptions involving lateral blasts.

Distinction of the morphological features of ideal dry avalanches can be impossible in ancient deposits. In addition, avalanches can display gradational features from deposits without juvenile eruptive material to deposits with juvenile material content comparable to those of pyroclastic flows. In the ancient rock record, therefore, avalanche deposits can only be distinguished with difficulty from other disorganized and poorly sorted deposits.
6.3 Deposition by glaciers

Glaciers are able to carry sediment as basal internal and surface load. Basal load is excavated and abraded from the substratum at the base and edges of the glacier. Some of it ends up within the interior of the glacier due to an upward component of the flow. Surface load results from melting of the glacier or from material falling or flowing onto the surface of the glacier, as for example, by rockfall, rock avalanches or eolian and pyroclastic air fall (Cas and Wright, 1987).

Glaciers deposit their loads at their bases, margins and termini in irregular piles or ridges called ground, lateral and terminal moraines, respectively. Deposits are characteristically composed of poorly sorted debris. Tills may have a compact and fissile texture due to lodgment under basal pressure. Melt-out in situ may lead to slightly sorted (clay- and silt-depleted) till. Debris may be reworked by various mass flows and glaciofluvial processes, forming better sorted outwash deposits.

The provenance of till depends on the source rocks in the drainage area of the glacial system. Tills having volcanic source rocks can be very similar to volcanic debris flows. Because vertical facies sequences can be complex, the recognition of glacial deposits can be very difficult. Especially in ancient deposits, if striations of boulders and substratum, frost-generated postdepositional structures and bullet-shaped boulders are not visible, the only reliable evidence might be the possible association with marine or lacustrine mudstones and shales containing dropstones.

6.4 Deposition by mass flows

Mass flows or sediment flows (Middleton and Hampton, 1973) or sediment gravity flows (Lowe, 1976, 1982) can be classified according to their dominant clast-supporting mechanism (Middleton and Hampton, 1973) combined with fluidal behavior (Nardin et al., 1979; Lowe, 1976, 1982). There are other classifications of different types of landslide or mass movement deposits (Johnson and Rodine, 1984) but many of them are based on morphological features and have few applications to internal structures.

The classification of Lowe (1976, 1982) presented in Table 3 gives a genetic explanation of structures in subaqueous mass flow deposits and is widely accepted and utilized with few modifications not only for subaqueous deposits (Stow, 1985, 1986;
Pickering et al., 1986) but also for subaerial mass flow deposits (Nemec and Steel, 1984). Deposition can take place by four mechanisms (Lowe, 1976, 1982). A mass flow or a part of it may freeze due to friction or cohesion. In fluid-rich mass flows or in fluid-rich parts segregated from avalanches or plastic mass flows, i.e. debris flows, clasts can accumulate by suspension and traction sedimentation. Suspension sedimentation is a fast and collective process. If the ability of a flow to carry larger clasts is gradually declining, larger clasts can start to settle towards the base of the flow leading to differential settling (Smith, 1986). Traction sedimentation takes place when clasts settle individually and the (traction) force induced by the current is no longer capable of dragging them back into motion (Lowe, 1982, Allen, 1984). Deposition from a mass flow can involve several of these mechanisms acting serially or simultaneously on different grain populations. Depositional processes are controlled by clast-support mechanisms. Therefore primary structures found in sediments reflect the support mechanisms just prior to deposition.

**Table 3** Classification of sedimentary gravity flows (Lowe, 1982).

<table>
<thead>
<tr>
<th>Flow behavior</th>
<th>Flow type</th>
<th>Sediment support mechanism</th>
</tr>
</thead>
<tbody>
<tr>
<td>fluid</td>
<td>fluidal flow</td>
<td>turbidity current</td>
</tr>
<tr>
<td></td>
<td>fluidized flow</td>
<td>escaping pore fluid (full support)</td>
</tr>
<tr>
<td></td>
<td>liquefied flow</td>
<td>escaping pore fluid (partial support)</td>
</tr>
<tr>
<td>plastic</td>
<td>debris flow</td>
<td>grainflow</td>
</tr>
<tr>
<td></td>
<td></td>
<td>mudflow or cohesive debris flow</td>
</tr>
</tbody>
</table>

The competence of a mass flow is a measure of its ability to carry a load. It is reported as the size of the biggest clast that can be moved (Pierson, 1981; Nemec and Steel, 1984). In mass flows, clasts can be supported by buoyancy, cohesion, turbulence, pore water pressure and dispersive pressure which results from collisions between particles (Bagnold, 1954; Hiscot and Middleton, 1979; Lowe, 1982). Several of these mechanisms are working simultaneously and flow conditions are strongly dependent on the active or dominant role they have. Within a flow, a grain-size population can be supported by a different mechanism than is another population, or different clast
support mechanisms can be working in different parts of the flow. Therefore, different mass flow types are only ideal end members, and there are complete gradations.

6.5 Deposition by debris flows (lahars)

6.5.1 Classification and competence of debris flows

Usually debris flows are considered to be processes by which granular solids admixed with entrained water, air and minor amounts of clay, move readily even on gentle slopes (Johnson and Rodine, 1984). However, there is not a generally accepted definition of a debris flow and the terminology used varies greatly. A debris flow after one classification can be an avalanche, a debris-slide, lahar or mudflow after another. Debris flows occurring in volcanic environments are commonly called lahars. Unfortunately, this term has been applied to describe any poorly sorted volcaniclastic deposit. In some works it has been used to describe "hot" deposits, some that are primary pyroclastic deposits. Because of its ambivalent meaning, Smith (1986) recommended that the term lahar be abandoned. However, the term has also been "redefined" as a general term for any type of a flowing mixture of a volcaniclastic material and water that is different from ordinary stream flows (Smith and Lowe, 1991). Because such a flow can go through complex changes in its rheological characteristics, a lahar is a volcano-hydrologic event and can refer to one or more discrete processes, but does not according to Smith and Lowe (1991) refer to a deposit. Changes of flow processes and rheological properties occurring during lahars are often referred to as flow transformations.

The, 1982 eruption of Mount St. Helens released a relatively dilute snow melt flood that incorporated volcaniclastic debris and became a debris flow (Pierson and Scott, 1985). Debris flows are also segregated from debris avalanches, forming a laterally juxtaposed apron in front of unconfined debris avalanche deposits (Scott, 1988; Palmer et al., 1991). It is somewhat unclear whether a pyroclastic flow can change to a debris flow by mixing with melt or stream water and condensation of vapor phase (Pierson, 1985; Smith and Lowe, 1991). Pyroclastic flow deposits are nevertheless, susceptible to reworking after coming to rest and are able to bulk up streams to become debris flows (Scott, 1985, 1988). Debris flows are generated by eruptions through crater lakes. Lakes may flood when crater walls also collapse or when a water body is
displaced by extrusion in the crater. Large debris flows and floods result from the
release of short-term lakes formed by lava flows or debris avalanches damming river
valleys. Debris flows originating in volcanic environments commonly have much larger
volumes (up to $10^7$ m$^3$) and mobilities than non-volcanic debris flows. Some of them
have been able to travel distances exceeding 100 km from their source areas within 3 to
9 hours (Naranjo et al., 1986). As a result of their complex origins, the clast
compositions of volcanic debris flows can be extremely variable. Debris flows resulting
from lake breakouts tend to have a heterolithic clast composition, with abundant
rounded clasts bulked from alluvial deposits and larger blocks representing natural dam
material. Debris flows involving mobilized pyroclastic material can be remarkably
homogeneous. They may also contain radially jointed clasts (the results of in situ
cooling), large crack-breciated blocks indicating origin from a debris avalanche (Smith
and Lowe, 1991). Because volcanic debris flows usually involve mobilization of
volcaniclastic material immediately or soon after eruptions, their detrital clay content is
lower than in debris flows occurring in non-volcanic environments. Most of the clay is
diagenetic. The proportion of clay- and silt-sized material has a great effect on
prevailing clast-support mechanisms.

Debris flows have been divided into cohesive debris flows and grain flows (Lowe,
1976, 1982). In cohesive debris flows, clasts are dominantly supported by cohesive
forces between clasts and the slurry mixture of water and finer sediments. The
deposition takes place as cohesion freezing when the shear stress in the flow no longer
exceeds the yield strength of the flow. In grain-flows of Lowe (1976, 1982) or
cohesionless flows of Nemec and Steel (1984), clasts are supported by dispersive
pressure created by collisions and deposited by frictional freezing (Middleton and
Hampton, 1973; Lowe, 1976, 1982). As the proportion of air in the interstitial fluid
increases, gravelly grain flows will gradually show characteristics of debris avalanches.
Clasts are commonly supported significantly by the buoyancy of a dense
sediment-water matrix. Where cohesion-free, they have been called density modified
grain-flows (Lowe, 1976, 1982).

The competence of debris flows and other types of mass flows is the sum of many
factors that are dependent upon cohesion, frictional strength and viscous resistance.
Many of these factors are interdependent. For example, the viscous factor becomes
important only when a debris flow (i.e., a plastic mass flow) loses its yield strength (i.e.,
it is no longer plastic). Cohesion is not dependent on flow thickness, but according to
theoretical concepts, the relationship between flow thickness and most other factors should be direct. Therefore the competence of an ideal debris flow can be considered to be a result of a constant cohesive factor and a non-cohesive resultant factor (Figure 2). Statistical studies of maximum clast size versus bed thickness plots (MCS/BTh-plots) may help to understand the relative proportions of cohesion and resultant non-cohesive factors. According to such studies, debris flows occurring in volcanic environments seem to have a larger scatter (Walton and Palmer, 1988) than non-volcanic alpine and alluvial debris flow conglomerates.

![Figure 2. Idealized model of the relationship between the competence and the bed thickness in debris flows (Nemec and Steel, 1984). The regression-line of the maximum clast-sizes and bed thicknesses measured from a single bed (or beds with similar lithology) at different locations. The intersection of the y-axis is defined by the cohesive strength of the flow. The slope of the line is controlled by the thickness-dependent factors.](image)

Matrix content or framework support is not itself a reliable criterion for distinguishing between cohesive or cohesionless debris flows. Both end members of debris flows can deposit totally matrix- or clast-supported gravels. The presence of an abundant mud matrix in pebbly sandstones or matrix-supported conglomerates; and breccias strongly suggests deposition by cohesive (though possibly turbulent) debris flow (Nemec and Steel, 1984). Muddy matrix can also be a factor in clast-supported...
conglomerates, it has been shown that as little as 5 percent muddy matrix can support bigger clasts by buoyant lift and can lubricate them to such an extent that the bulk strength is practically controlled by the cohesive strength of the muddy matrix (Johnsson and Rodine, 1984). Unfortunately, in ancient deposits the clay content cannot be easily estimated; it is difficult to distinguish the primary detrital matrix from the diagenetic matrix or postdepositional hydrothermal alteration products.

In spite of the terminological and the theoretical controversy, there is a general consensus that debris flows show plastic behavior and that they are deposited by freezing en masse. Most studies of debris flows are made from non-volcanic alpine or alluvial flows in semiarid climates. Many of these flows result from gravitational instability of regolith or weathered bedrock. According to these studies, the plasticity of debris flows follows the so-called Coulomb-viscous model (Johnsson, 1970; Johnsson and Rodine, 1984; Nemec and Steel, 1984). Although shearing is strong in the bottom of debris flows, only a few of them show signs of basal sliding. Therefore, they are not solely "plug flows", but have a basal zone of laminar flow. Actually a debris flow freezes when the rigid or plug part of the flow reaches the bottom (Johnson and Rodine, 1984). In addition to laminar and plug flow behavior, parts of some debris flows seem to have been turbulent (Lowe, 1976, 1982; Nemec and Steel, 1984).

6.5.2 Characteristics of debris flow deposits

Non-volcanic debris flows usually have sheet-like to highly lenticular geometries (Nemec and Steel, 1984) although in their proximal parts, debris flows can be restricted to channels (Johnson and Rodine, 1984). Basal erosion is typically insignificant or limited to typically proximal slopes where debris flows are accelerating. Beds do not usually show internal stratification, although some surging debris flows may be crudely layered. Successions of subsequent flows may be well-bedded because of distinct bed boundaries. The textures of beds vary from polymodal to bimodal and from clast- to matrix-supported. Beds also commonly contain some outsized cobbles or boulders.

Volcanic debris flows usually have much larger volumes and their mobility is higher than in nonvolcanic debris flows. Their deposits are typically thinner than the flows themselves. Deposits can form thin veneers of matrix material with lags of coarse clasts. Instead of smoothing, these veneer deposits tend to mantle existing topography.
Recent volcanic debris flows form depositional terraces (Rodolfo, 1989) and clast-supported, inversely to normally graded streamlined bedforms described as "whaleback bars" (Scott, 1988). These bedforms are formed by basal divisions of debris flows associated with volcanism, as at Mount St. Helens. Basal carpets of cobbles and boulders were formed during rising and peak discharges of debris flows. At medial distances these concentration carpets move as waves or pulses, the sound of which can be heard by witnesses. These pulses deposit coarse clast-supported gravel bars with the crests nearly reaching the peak flood-level. In more proximal parts, bedforms are not developed and gravelly channel fill layers are deposited. Gravel is mainly derived from underlying gravelly alluvium. The major part of a debris flow flood wave consists of finer (pebbly) gravel dispersed in the flow or buoyed at the flow surface. Their deposits are matrix-supported gravels and conglomerates covering flood-plain areas. The matrix is characteristically granular sand with very little mud. Recent and ancient debris flows around Mount St. Helens have abraded bases or show distinct basal sandy structures called sole layers (Scott, 1988). At the abraded surfaces, the matrix of underlying deposits is not polished or grooved. However, boulders and cobbles are truncated by grinding and impacts by boulders in basal flow. Apparently, the overlying flow induces pressure that holds boulders in the unconsolidated substrate while they are being truncated.

The material of the sole layers and the matrix of overlying deposits are similar. The contact of sole layers with overlying deposits are commonly sharp but can in places appear gradual over several centimeters or poorly distinguishable. Scott (1988) distinguishes three types of sole layers. (1) Massive sandy sole layers lacking dispersed pebbles are common in floodplain debris flows. The sole layers are rarely compacted. Typically they are a few decimeters thick. (2) The sole layers with dispersed pebbles are found at the bases of coarse channel fill layers and bedforms. The pebbles giving characteristic bimodal distribution are rarely greater than 32 mm in diameter. Grading patterns are variable but inverse grading, at least in the lower part of the layer, is most common. Locally, the pebble-dispersed sole layer can be crudely stratified with few textural variations between strata. Locally stratification is apparent from slight changes in color. Color changes reflect a lateral zoning in the composition of the overlying coarse deposits. Sole layers can be more compact and resistant to erosion than overlying deposits. They may also show primary bed-parallel foliation. Pebbly sole layers are typically 10 to 50 cm thick and consist of 10-15 percent of the total thickness.
of a flow. (3) The third type of sole layer is found at the channel bases of huge ancient debris flows. The layers are clast-supported, well-rounded pebbly conglomerates in which the matrix consists of silty-sand. Upper contacts are usually sharp but can be gradual over a few centimeters. The sole layer is 0.5 -1.1 m thick in a depositional unit that is 6-7.5 m thick (Scott, 1988).

The migration of the laminar-plug flow boundary and the distribution of shear in the flow has affected the formation of sole layers (Scott, 1988). In places, the sole layer develops above easily erodible sediments which do not have an effect on their character. Their material is probably not derived from underlying deposits but instead, their formation is related to basal shearing and the formation of inverse grading. Sole layers lacking pebbles have been found in floodplain deposits. Such layers have probably been formed by passive flows. Layers do not develop in floodplain deposits that overlie rough surfaces on which the flow was slow. Dispersed pebble-containing layers are associated with channeled flows. These sole layers are formed during or before the flow peak by accretion over a significant time interval. The formation of the sole layer is favored by the lack of roughness on the underlying surface, and the intensity and duration of basal shearing. Once formed, a sole layer is not easily eroded. A sole layer can become intensely compacted and sheared. A locally sharp upper boundary is formed when an interval of sustained shear compacts the layer and interrupts the process that otherwise would form an inversely graded basal zone. Clast-supported gravelly sole layers with well-rounded clasts are formed under the greatest intergranular stresses in the large debris flows. Abraded surfaces are formed in places where changes in depth and velocity cause the boundary between laminar and plug flow to approach the base of the flow. This causes concentration of shear and localized abrasion.

The volcanic debris flow deposits studied by Walton and Palmer (1988) deviate from the non-volcanic deposits studied by Nemec and Steel (1984) by showing larger scatter in MSC/BTh-plots (maximum clast-size vs. bed thickness) described above. According to Smith and Lowe (1991), this indicates that the thickness of the volcanic debris flow deposits does not necessarily correlate with the transportation competence and the thickness of the moving debris flow itself. The shear resistance and competence of volcanic debris flows are derived from grain collisions that are not accounted for by the Coulomb-viscous model and cohesive forces of silt-sized fine ash particles. Altogether these characteristics suggest that volcanic debris flows deviate from ideal cohesive
debris flows and their deposition does not necessarily take place *en masse*. However, the volcanic debris flow deposits are very similar to non-volcanic alluvial and alpine debris flows and can be distinguished based on similar criteria (Smith and Lowe, 1991).

Beds can be ungraded to well graded depending on the internal regime (Schultz, 1984). Ungraded debris flows or parts of the flows represent plug flow in which shear strength is high due to cohesion, with additional friction and viscosity preventing turbulence and effective clast collisions (Johnson and Rodine, 1984; Nemec and Steel, 1984). Such flows are typically massive and matrix-supported. Inverse grading may develop in a thin basal zone (Shultz, 1984, flow regime I). In clast-rich debris flows, friction is dominant over cohesion. Such flows are typically clast-supported, and inverse grading may be well-developed but limited to the thin basal parts (flow regime II, Shultz, 1984). In more fluidal flows, turbulence becomes important. Deposits are clast-supported and commonly inversely-to-normally graded. However, non-graded and normally graded beds are not uncommon (flow regime III, Shultz, 1984). In extremely fluidal flows, coarser material may segregate into a bedload layer in which flow takes place in a grain-flow fashion (flow regime IV, Shultz, 1984).

For fluidal mass flows, it has been suggested that nongraded flows are attributed to a balance between dispersive pressure and gravitation (Lowe, 1979, 1982). Typically the deposition of inversely graded gravels is not preceded by deposition of normally graded deposits. This suggests that the sensitivity mechanism proposed by Naylor (1981) is insignificant for the origin of the inverse grading (Walker, 1984; Scott, 1985, 1988; Walton and Palmer, 1988). Inverse grading is usually attributed to dispersive pressure resulting from basal shear. In deposits representing flows transitional between debris flows and more fluidal run-out deposits, inverse grading may develop throughout the thickness of the bed. This is explained as the product of the progressive deposition of the hyperconcentrated part of the flood followed by a gradually thinning debris flow (Scott, 1985, 1988). In debris flows, changes in slope or roughness of the basal surface can cause the boundary of the rigid plug to migrate, producing shear and inverse grading in various parts of a flow.

Normal grading probably indicates a considerable role of turbulence (Lowe, 1982; Nemec and Steel, 1984). In fluid-rich flows, normal coarse-tail grading can be produced by collective, instant deposition of the suspension load, or by frictional freezing of a graded flow (Lowe, 1982). In debris flows, normal coarse-tail grading is a
result of the differential settling of coarser clasts that cannot be supported by plastoviscous flow (Scott, 1988).

Debris flows having a polymictic clast composition usually have a homogenous matrix and a uniform vertical distribution of different clast types. As noted by Walton and Palmer (1988), none of the attributed support mechanisms can explain the good mixing of source materials. According to them, the explanation of a homogeneous matrix could be the bulking of a turbulent surge or flood into a debris flow with a rigid plug, as occurred at Mt. St. Helens (Scott, 1985, 1988).

6.6 Deposition by fluidal mass flows

Debris flows may contain more than 80 percent sediment. As the fluid content increases, the flow characteristics change from plastic to viscous. Turbulence and pore water pressure become important support mechanisms. Submarine viscous mass flows have been described as liquefied, fluidized or turbidity (density) flows, depending upon their dominant clast-support mechanism (Middleton and Hampton, 1973; Lowe, 1982). Some turbulent mass flows show imbrication in which the long axes of pebbles are gently inclined down towards the upstream direction. Such an orientation is not likely a product of rolling.

Fast fluidal flows found in subaerial environments are transitional to heavily-laden stream flows. Fluidal flows are found at the termini of glaciers (Lawson, 1982) and alluvial fans where they can triggered by heavy rainfall (Pierson, 1981). In a volcanic setting, fluidal mass flow deposits commonly underlie debris flow deposits. Such deposits represent high-concentration flows that were formed from debris flows by dilution and eventually bypassed slower moving debris flows. Dilution may be due to added rain, stream or melt water. An opposite flow transformation from fluidal mass flow or stream flow to a more concentrated flow type may take place if water infiltrates the underlying permeable volcanioclastic deposits. The internal structures of such flows are not well-described. They seem to deposit sheet-like bodies which vary from slightly channeled and weakly graded to strongly channeled and normally graded or even thoroughly stratified (Pierson, 1981; Lawson, 1982). Volcaniclastic deposits showing characteristics transitional to debris flows and dilute fully turbulent stream flows are often referred to as hyperconcentrated flows (Smith, 1986; Walton and Palmer, 1988; Smith and Lowe, 1991). The importance of hyperconcentrated flow deposits in modern
and ancient volcaniclastic sequences has been recognized only relatively recently; the first papers concerning their deposition were published in the 1980's. Gravel-sized hyperconcentrated flow deposits described by Smith (1986, 1988) are poorly sorted and clast-supported. They commonly show a normal distribution grading, and generally lack a basal inverse grading. These deposits can be very similar to high-velocity streams and floods. However, hyperconcentrated flow deposits lack stratified sand beds and lenses found in many fluvial gravel deposits. In clast-supported stream conglomerates, open spaces between framework cobbles and pebbles are often filled with distinctively finer sandy matrix by infiltration after the deposition of the coarse clasts. In hyperconcentrated flow deposits, the framework spaces are often filled with coarser clasts that probably have not infiltrated. This in turn, suggests rapid deposition of a wide grain-size range of sediments. Clast orientation is commonly bimodal. Large boulders and cobbles tend to have their long axes perpendicular to flow direction, whereas pebbles are oriented with long axes parallel. Such imbrication implies that larger grains were transported by traction load while smaller grains were in turbulent suspension (Smith, 1986; Harms et al., 1984).

Sandy deposits described by (Smith, 1986) lack cross-bedding but commonly are horizontally stratified. Strata vary from a few mm to several centimeters in thickness, being on the average very thin beds according to Ingram's (1954) classification. Stratification is defined by the alternation of coarser and finer layers, with the overall medium-grain size close to coarse sand. Because the coarser layers also contain finer material, the layering is not sharp. Stratification is usually more distinctive in pumiceous deposits. Layers can usually be traced laterally for 2 to 10 m.

Single depositional units are usually several meters thick, but deposits more than 30 m thick without any internal erosion surfaces have been found (Smith, 1986). Deposits commonly contain over-sized erosion surfaces have been found (Smith, 1986). Deposits contain over-sized clasts up to 1 m in diameter, probably transported by rolling or sliding in bedload. In dilute upper flow regime stream flows, lamination can be formed by slight fluctuations in the basal boundary layer (Allen, 1984). To explain less distinct stratification of hyperconcentrated flows, Smith (1986) called upon migrating low-amplitude, long-wave length, dune-like bed forms.

Some of the deposits have massive, normally graded bases which change gradually upward to horizontally stratified parts. The lower parts can be coarse gravel or sand. The upper stratified part is usually medium- to coarse-sand with lenses or patches of
dispersed or clast-supported pebbles (Smith, 1986). Such deposits are similar to graded-stratified high-density turbidity of fluidized flow deposits found in subaqueous settings.

Recent deposits described by (Scott, 1988) consist of sand or pebbly gravel. Beds are clast-supported and massive or rarely stratified. If coarse-tail normal grading occurs, it generally involves the concentration of pumiceous fragments toward the bases of the deposits. Deposits may contain inversely graded subunits, as for example the laharrunout facies of Scott (1988).

Hyperconcentrated flow deposits have a wide variety of lithofacies characteristics which suggest that clast-support is mainly provided by turbulence or escaping pore fluids. A variable proportion of the sediment load is transported by tractional processes. The resulting deposits show marked analogies to subaqueous mass flow deposits. The "real" structures and lithofacies sequences observed in subaqueous mass flow deposits (e.g. Pickering et al., 1986) seldom strictly follow simplified "ideal" depositional models (Lowe, 1982). Similarly, the differences between the deposits described by Smith (1986), Walton and Palmer (1988) and Scott (1988) probably do not reflect differing angles of scientific approach, but instead reflect a diversity in depositional characteristics of hyperconcentrated flows.

6.7 Deposition by fluvial processes

If sediment content is less than 40 percent, the clast interactions are small and clasts are moved grain-by-grain (i.e., by traction transportation). Traction transportation can take place by sliding, rolling or jumping. Bedload transportation refers to sliding and rolling. Jumping results in short-term suspension known as saltation. Fine sediments are also transported by suspension. Suspension transportation of sand can be important during peaks of flows or flood events. In non-volcanic alluvial environments, gravelly fluvial sediments are almost completely braided stream deposits. These deposits are typically clast-supported and massive or crudely to well-stratified (Miall, 1978, 1984; Rust, 1978; Harms et al., 1984; Nemec and Steel, 1984; Rust and Koster, 1984). The lithofacies sequences are very variable. Gravels deposited by flash flooding, particularly alluvial fan stream deposits, tend to have the greatest textural immaturity. The textural characteristics of such deposits may resemble those of primary debris flows and distinguishing of such deposits from clast-supported hyperconcentrated or debris flow
gravels may be difficult. Possible evidence for fluvial origin includes scouring or channeling, widespread imbrication, and vague to distinct stratification. However, distinguishing of beds can be difficult, especially in texturally immature flash flood deposits. More mature texture suggests stronger channeling, more continuous runoff and effective contemporaneous reworking (Nemec and Steel, 1984). Cobbles and pebbles transported by rolling on the bottom commonly show imbrication, with the long axes transverse to the channel and intermediate axes dipping upstream (Harms et al., 1984).

The thickness of strata and cross-stratified sets may provide a rough indication of water depth. Strata only a few centimeters thick result from rapidly shifting shallow flow at the tops of braid bars or gravel sheets, or from unchanneled flooding on the lower reaches of alluvial fans. Thicker strata suggest deeper channeled flows as in primary fan channels or strata channeling flood events (Nemec and Steel, 1984). Well-developed stratification with alternating sandy and gravelly layers can be created by slight fluctuations in stream velocity (Harms et al., 1984).

Typical gravelly lithofacies found in braided streams are massive to crudely bedded and show horizontal stratification and imbrication. These can be interpreted to represent longitudinal bars, lag, and/or sieve deposits. Trough cross-bedded gravels are probably minor channel fillings. Planar cross-bedded gravels can be interpreted to represent linguoid bars or deltaic growths from older bar remnants (Miall, 1978, 1984). There does not seem to be a simple facies model explaining all the braided systems (Harms, 1984; Collinson, 1986) and it has been suggested that fluvial systems would be better outlined by the recognition of architectural elements from large three-dimensional outcrops (Miall, 1985). However, certainly the deposition of either horizontally or cross-stratified braided stream sediments takes place under waning flow conditions or declining discharge. A common sequence consists of a composite unit of upward-finining conglomerate usually a few decimeters to a few meters thick, capped by sandstone. When upward-finining is well-developed, such sequences may be very similar to normally graded gravelly units resulting from almost instantaneous sedimentation from fluidal mass flows (e.g., Fig. 9 of Nemec and Steel, 1984).
7. Transportation processes and sedimentation

Generally, reworking and recycling sedimentation processes are important but variable in volcanic terrains. The magnitude of sedimentary processes may be greater in volcanic than in non-volcanic terrains. Some volcaniclastic deposits remain soft and poorly consolidated or their strength may be weakened by hydrothermal alteration and weathering. Abundant volcaniclastic debris is very susceptible to erosion by heavy rain or meltwaters of glaciers. Huge piles of volcaniclastic deposits are susceptible to slides, free fall, avalanches and debris flows. On the average, the mobility of avalanches recorded in volcanic environments is higher than in non-volcanic environments. This is explained by several factors. Alternating soft pyroclastic and rigid lava or welded tuff layers have a high structural instability. Parts of volcaniclastic deposits may be strongly hydrothermally altered, especially around eruption vents. The growth of cryptodomes and other shallow intrusions can cause fractures prior to initiation of sliding. Heat discharge from shallow intrusions may heat ground water to the supercritical state. The sudden release of pressure may cause boiling of supercritical water, producing additional fracturing in the volcanic edifice. Also, the relief of a volcanic edifice may be substantial.

Glaciers may provide an important contribution to the volcaniclastic depositional record. Although most of them are located at high latitudes, glaciers are found at high altitudes, even in the equatorial zone (e.g., Mt. Kilimanjaro). If the glaciers are small, glacial deposits do not contribute much to the ancient rock record. However, in some volcanic areas, such as in Iceland, glaciers are major epiclastic agents. The rapid melting of ice may cause lahars or floods. Floods can be as huge, as are "jokulhaups" in Iceland, in which the discharge exceeds that of the Amazon River.

Floods may result from heavy rainfall. In the streams on the slopes of Volcano Fuego in Guatemala, peaks of floods induced by rainfall are able to carry 2-3 m boulders as bedload.

The Volcano Fuego is also a good example of the temporal relationship between volcanism and sedimentary processes. During an inter-eruption phase, which may last more than a century, sedimentation rates are low, meandering rivers are incised and deltas are reworked. Eruptive phases usually last less than a year, during which huge amounts of lavas and pyroclastic falls and flows are deposited. Subsequent to eruptions,
debris flows and coarse fluvial sedimentation occur and within 1-2 years alluvial fans are built around the margins and bases of stratovolcanoes. The fan-building phase is followed by the braiding phase that lasts from 20 to 30 years. During this phase, incised meandering rivers turn into flood-prone braided systems and deltas prograde rapidly. In other parts of the active margin in Central America, increased sediment supply following eruptions creates elongate valley pond lakes, reactivates older alluvial fans, and causes significant building of deltas.

Subaqueous sedimentary processes are not well studied in modern volcanic areas. Marine depocenters derive a large proportion of their volcaniclastic debris from subaerial and shallow marine parts of volcanic systems. Studies of ancient systems suggest that in wholly subaqueous volcanic systems, various gravitational collapses are the main transportation processes. In restricted basins, extra-formational sediment transportation from shallower basin margins can be important (Cas and Wright, 1987).

8. Volcaniclastic deposits and facies analysis

The environment of deposition usually cannot be defined on the basis of a single depositional structure or texture. The physical conditions that control the formation of a depositional structure can be similar in several environments. The deposits reflect physical conditions prevailing at a depositional environment. Therefore, the changes in the vertical and horizontal distribution of the deposits are better indications of environments. The definition of the environments of deposition, ancient or recent, is a process that starts with observations on the deposits. The interpretation is done after the changes of depositional structures and textures, vertically and horizontally, have been recognized. The process is called facies analysis. A facies is a body of a rock that shows a set of objectively established characters that separate it from the remaining parts. If the characters are all lithological or biological, the terms lithofacies or biofacies can be used. Anderton (1985) defines an interpretative facies as a part of the depositional record that shows a set of structures that indicate an environment of deposition (e.g. a tidal facies or a fluvial facies). Commonly an interpretative facies itself is a set of descriptive facies (Anderton, 1985). The term lithofacies has also been used with a similar meaning to Anderton's (1985) interpretative facies (Bates and Jackson, 1984; Anderton, 1985).
Facies interpretation explains how facies were formed and defines the interactions of facies. Commonly this is done after visual observations, but also statistical methods have been used (Harper, 1984, Walker, 1984, Reading, 1986). Attention must be paid to facies boundaries, series, cycles and assemblages. The understanding of the interactions between facies is a key to the facies model. The facies model explains the physical conditions and their changes during the deposition. A model can be formulated for a single depositional unit (for example a Bouma bed, a braided river or a prograding delta). Depending somewhat on one's philosophy of life, the the facies models can have a variable purpose and significance (Anderton, 1985). According to Walker (1984), a facies model is a sequence of facies that provide

- a norm that defines what is typical or atypical for the deposits,
- a framework for the studies of facies changes in different areas,
- a prediction in other geological environments,
- a summary on which the interpretation of a system or an environment of deposition is based.

According to Anderton (1985), a facies model is a description or interpretation of environments of deposition or a part of them, their origin, their characters and their changes that is based on realistically defined attributes. For Anderton (1985), every model is more or less unique, although there are similarities between the models of similar environments. Miall (1985) also emphasized the continuous spectrum of facies models and established a new facies analytical approach that is based on "architectural elements". However, the method is poorly applicable to two dimensional exposures.

The texture and fabric of reworked volcaniclastic deposits (secondary pyroclastic deposits of Lajoie, 1984) and recycled (epiclastic) deposits may be very similar to primary pyroclastic debris so their distinction can be difficult. The depositional mode of various coarse and poorly sorted volcaniclastic debris is difficult to distinguish, even if there is no alteration or weathering of unstable vitric fragments to clay, zeolites and carbonates. Tills and lahars may be distinguished by studying deformational microstructures by scanning electron microscopy (Mahaney et al., 1991) but this method may not be applicable to rocks which have had diagenetic or metamorphic changes in their texture. In many cases the only evidence of a reworked or redeposited origin is a polymictic pebble and cobble content and the depositional mode must be
only assumed from lithofacies sequences and associations. Unfortunately, not many large-scale facies studies of volcaniclastic deposits have been completed and facies models for volcaniclastic rocks either are tentative or "in preparation" (Suthren, 1985; Cas and Wright, 1987). In the cases when the source has been monomictic, epiclastic deposits may have textures that are very similar to the original source. Therefore, the contribution of epiclastic processes in the ancient rock record is probably underestimated.
Studied deposits of the Keetley Volcanics

In stratigraphic studies, deposits usually are approached with emphasis on the origin of defined and correlated lithological units (Miall, 1984, p. 5). Lithostratigraphic units are divided in declining hierarchy into supergroups, groups, formations, members and beds. Terms like tongues or wedges are sometimes used to describe interfingering members. Units resulting from a single volcanic depositional event are sometimes referred to as flows instead of beds (Miall, 1984, p. 78-79). Formations are the principal map units of traditional stratigraphic studies. Definitions of stratigraphic units should be done after establishment and interpretation of the framework of correlated sequences (Miall, 1984, p. 5). Thus far the Keetley Volcanics have not been divided into traditional stratigraphic units and it would be tempting to subdivide the Keetley Volcanics into members and beds. However, the few studied stratigraphic sections are not an adequate basis for lateral correlation. Therefore, the studied deposits are not given stratigraphic names in this report, but the descriptions should be useful in subsequent naming of subunits within the deposits.

Each lithology described is also a descriptive lithofacies and interpretations of their mutual relationships, origins and environments of deposition can be done. If one prefers the view of Anderton (1985), the following is a facies analysis of the Keetley Volcanics. However, the few studied deposits of the Keetley Volcanics do not provide a good basis for a facies model as outlined by Walker (1984). Taking also into account that facies analysis of volcaniclastic deposits is largely in progress, the facies terminology was not used.
9. Petrography

9.1 Breccia clast mineralogy and magma source/sources

Breccia clasts give direct information on the magmatic source or sources of the Keetley Volcanics. Except for a few samples, all the studied sparsely vesicular volcanic breccia clasts are porphyritic andesites. Other sparsely vesicular clasts are basaltic-andesite or dacite based on nature of phenocrysts. Modes (based on 600 points) of breccia clasts and porphyritic intrusive rocks are presented in Table 4. Major components include plagioclase, amphibole and groundmass. A few samples include clinopyroxene or biotite as major minerals, but typically these mafic minerals as well as orthopyroxene are present only in minor amounts. Other uncommon accessory minerals are apatite, opaques, quartz, sanidine, sphene, and zircon. The groundmass shows varying degrees of alteration and devitrification. Alteration products include zeolites and chlorite-smectite. Most of the breccia clasts seem to have a relatively glassy groundmass with 20-30 percent small feldspar laths. Also, microcrystalline and completely glassy clasts can be found. Some of the feldspar microlites have a very low relief and staining for K-bearing minerals indicated that such microlites are probably sanidine. Plagioclase crystals are subhedral or euhedral, compositions vary from oligoclase to andesine, although albite-rich overgrowth rims can be found. Plagioclase phenocrysts may have rims and cores consisting of irregular intergrowths of K-feldspar or glass blebs. Such sieve-textured rims and cores are typically found in larger phenocrysts. In the samples containing abundant sieve-textured phenocrysts, plagioclase crystals commonly have slightly rounded edges and embayments are common.

Amphibole phenocrysts are typically compositionally zoned oxyhornblende (pleochroic colors from light yellowish-green (x') to dark rusty brown (z'), 0-50° extinction angle) and many have thin opaque oxidation rims around them. In a few samples, the amphibole is green and distinctly in the hornblende range.

Biotite is commonly strongly oxidized and may be partly replaced by opaques. Some of the biotite clasts have small rounded glass inclusions. The clinopyroxene present is augite, based on its optical properties. One sample has compositionally zoned phenocrysts in which the small 2V-angle of the core suggests a pigeonitic composition.
Some clinopyroxene phenocrysts have amphibole and even biotite inclusions. Hypersthene is typically found as small crystals (Ø 0.1 - 0.5 mm).

Breccias containing abundant dacitic clasts are found at Silver Creek Canyon (near Interstate Highway 80, Appendix 4). The matrix of a debris flow at outcrop 8 contains abundant dacitic rock fragments and volcanic quartz grains. Also a sample from outcrop 12 (Appendix 4) indicates a dacitic source component for at least some of the distinctively polymictic debris flows of that study area. A few pebbly volcanic sandstones consist of dacitic sand- and gravel-sized clasts (Jordanelle area, roadcuts 2 and 3). These deposits will be discussed in the following section. Accessory quartz crystals are present at outcrop 6 near Jordanelle dam (Appendix 2). These accessory quartz crystals (Table 4, samples JOROC6 2a and JOROC6 2b) have rounded shapes, corroded edges and embayments similar to the above mentioned deposits. However, atypical of volcanic rocks, the quartz crystals show undulatory extinction. Since the samples lack microscopic structures suggesting brittle or plastic deformation, the quartz crystals are probably resorbed xenocrysts from acid wall rocks rather than postdepositionally strained phenocrysts.

Basaltic-andesitic breccia clasts are clinopyroxene-plagioclase porphyritic and contain small feldspar laths in a glassy or aphanitic groundmass. In spite of the more mafic mineralogy, the weak but distinctive staining of the groundmass indicates that these rocks contain K and therefore possibly approach a shoshonitic chemical composition. Scoria breccias, lava flows and polymictic volcanic breccias containing basaltic-andesitic rock fragments seem to be lacking in lower stratigraphic levels.

Breccia clasts include common grayish-white to light gray clasts that are strongly altered to clay. Clast sizes vary from blocks to granules. These clasts are not derived from a single rock type, but because of the strong alteration, different sources could be identified in the field only with difficulty. Microscopic studies indicate that the clasts include pumice, especially in the volcanic sandstones (described below) which contain dacitic rock fragments. Megasopic observation also indicated that some of the strongly altered clasts are alloclasts of pumiceous tuff. Such altered clasts are fragments of a pumiceous lapilli tuff or breccia. Some of the samples were highly vesicular but the texture is scoriaceous rather than pumiceous. Some samples contain plagioclase, amphibole, biotite phenocryst and quartz xenocrysts. Their andesitic mineralogical composition is similar to the sparsely vesicular clasts present at the same outcrop.
Table 4. Point count modes (600 points) of breccia clasts.

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Sample coding
JOROC8 4  Jordanelle area, outcrop 8, sample 4, location in Appendix 1
JORRC1 A  Jordanelle area, roadcut 1, sample A, location in Appendix 1
H10 2B     Indian Hollow, outcrop 10, sample 2B, location in Appendix 2
IH INTR, JORD INT etc. Intrusive rocks from Indian Hollow, Jordanelle, Francis and Provo River Valley, locations in Figure 1
Without statistical analysis, it seems that the distribution of plagioclase and amphibole contents show opposite trends and possibly follow a lognormal distribution (Figures 3 and 4). Variations of phenocryst contents could indicate that breccia clasts were derived from several volcanic rock types. However, the variation could suggest a source with a heterogeneous phenocryst content such as a vertically and compositionally zoned magma chamber which would be consistent with the geochemical results of Woodfill (1972). All the other phenocrysts were present in minor or accessory amounts. A few samples of dikes cutting andesitic breccias or scoria breccias associated with lava flows have a basaltic-andesitic composition. The samples of andesitic volcanic breccia clasts have a relatively uniform mineralogy with variations only in the accessory phenocrysts.

Petrographic evidence of magma mixing such as ocelli consisting of amphibole or clinopyroxene rims around quartz crystals or coexisting unstable high- and low-temperature mineral pairs (olivine/quartz or Ca-rich clinopyroxene/hypersthene) were not found (Gangs et al, 1988; Hibbar, 1991). Such textures would indicate a distinctive drop in the temperature and a transient thermodynamic disequilibrium during the crystallization. Rapid changes can be expected if mafic and felsic magmas (with characteristically lower temperatures than mafic magmas) are mixed. Sieve-textured plagioclase and the replacement of amphibole by groundmass (Figures 5 and 6) indicate resorption. Resorption has probably occurred due to convective "mixing" and other processes occurring in large magma chambers. Mafic enclaves consisting of aggregates of plagioclase, clinopyroxene, amphibole and alteration products such as opaques and rusty or greenish mesh (iddingsite and chlorite) are common. Therefore, the observed enclaves may represent either xenoliths or blobs of mafic coeval magma.

An example of ocellar texture is shown in Figure 7. A core consisting of opaque, chlorite, and plagioclase is surrounded by a light-colored amphibole (actinolite?) which gradually changes to green amphibole (hornblende?). The core may represent a ferromagnesian (xenolithic or cognate) inclusion. The reaction rim resulted from a chemical instability of the ferromagnesian core but does not indicate a distinctive change of temperature during crystallization.

The observed enclaves, ocellar structures and (biotite?) inclusions within clinopyroxene do not confirm rapidly descending crystallization temperatures and thermodynamic disequilibrium conditions commonly resulting from magma mixing.
However, their abundance together with the common evidence of resorption suggests magma/country rock interactions and/or complex crystallization (Didier and Barbarin, 1991; Hibbar 1991). Most of the studied clasts are andesites and show slight variation in accessory minerals. The comparison of the samples from different study areas is unreliable and could reflect biased sampling. However, based on mineralogy only, the sampled shallow intrusions and dikes could have provided sources for the thick breccias of the Keetley Volcanics.
Figure 3. Distribution of amphibole phenocryst content in the studied breccia clasts (53 samples).

Figure 4. Distribution of plagioclase phenocrysts in studied breccia clast samples.
Figure 5. Photomicrograph of the granular matrix between the basaltic-andesitic breccia clasts from outcrop 21 at Jordanelle area (Appendix 1). The rock consists almost entirely of clinopyroxene-amphibole-plagioclase porphyritic basaltic-andesitic clasts. Amphibole phenocrysts have opaque rims and have been partly replaced by the groundmass. Clinopyroxene phenocrysts have tiny opaque and amphibole inclusions. Subparallel nicols. The large partly replaced oxyhornblende phenocryst is 4 mm long.
Figure 6. Photomicrograph of a volcanic breccia from outcrop 8 of Jordanelle area (Appendix 1). The large andesite clast forming the lighter colored part contains abundant plagioclase phenocrysts with sieve-textured cores and rims. The darker colored part consists of breccia matrix. Subparallel nicols. The largest plagioclase crystals are 2 mm long.
Figure 7. Photomicrograph of an ocellar structure in a dacitic pumice fragment. (Jordanelle area, roadcut 3, Appendix 1). The altered opaque, chlorite and plagioclase-containing core is surrounded by the light-colored rim of amphibole (actinolite?) that gradually changes to green amphibole (hornblende?). The ocellus is 2.5 mm in diameter. The blue color is due to tinted epoxy.
9.2 Clast composition and the depositional processes

Redeposited volcaniclastic deposits may be very similar to primary pyroclastics and their interpretation is commonly based on the distinction between a polymictic or a monomictic clast composition, and the study of depositional structures (Cas and Wright, 1987). In many beds of the Keetley Volcanics, the distinction was difficult; because of the narrow mineralogical variation, different andesite lithologies were recognized with difficulty.

Difficulties are partly due to the porphyritic texture of the studied volcanics. A porphyritic volcanic rock which contains plagioclase, amphibole, pyroxene and biotite phenocrysts can be broken into granules or sand-sized fragments that lack phenocrysts or show any combination of one or more phenocryst-minerals. Comparison of clast types is also hampered by alteration, oxidation and devitrification. An example of the effect of oxidation is shown in Figure 8. The sample consists of well-rounded granule-sized clasts which first appear very different from each other. However, the sample is taken from a horizontally stratified basaltic-andesitic volcanic breccia/sandstone, 1 m thick that is between two pyroxene-amphibole porphyritic lava flows. The lava flows and the breccia deposits have a basaltic-andesitic mineralogy and therefore, are distinctively different from the underlying and subsequent andesitic volcaniclastic deposits. The lava flows have red oxidation surfaces and contain minor agglutinated spatter. The basaltic-andesitic breccia/sandstone sequence between the lava flows contains abundant rounded amygdaloidal basaltic andesitic clasts that also show a variable degree of oxidation; they vary in color from black to rusty brown. Under a microscope, the glassy groundmass of these clasts varies from black (strongly oxidized) to light grayish-brown. The vesicles are filled by brownish zeolite which conceals the amygdaloidal scoriaceous texture in Figure 8. The plagioclase crystals are small, less than 0.15 mm in length and there seems to be a gradation in size from the tiniest to the largest plagioclase laths. However, the sample contains ca. 5 percent light gray clasts in which the groundmass contains more microcrystalline, commonly quenched tiny plagioclase laths and in which small plagioclase crystals (Ø 0.1 - 0.2 mm) do not show a continuous size-distribution.
Figure 8. Photomicrograph of granular basaltic-andesitic volcanic breccia/sandstone (outcrop 21, Jordanelle, Appendix 1). Largest grains are approximately 1 cm in diameter. Subparallel nicols.
The well-rounded basaltic-andesitic clasts probably are cinders that were formed during the same volcanic period as the mineralogically similar basaltic-andesitic lava flows and which were variably oxidized. The clast composition of these deposits is much more monomictic than it first appears, and the good rounding of the clasts may reflect the eruption style, rather than long transportation. However, the accessory clasts and distinctive stratification suggest that the scoriaceous material was reworked and partially mixed with debris from another source.

Examples of the effect of devitrification and/or alteration are shown in Figures 9 and 10. At Indian Hollow, deposits include breccias that contain plagioclase-amphibole porphyritic clasts. The dominant clasts have a bluish-black glassy matrix. Clasts also include minor pumice clasts and plagioclase-amphibole porphyritic clasts with an aphanitic groundmass and accessory clasts with a reddish-brown aphanitic glassy groundmass. The modes of different samples are given in Table 4 (Samples IH3 BLACK, IH3 GRAY and IH3 RED). Especially the grayish clasts with aphanitic groundmass can easily be misinterpreted to be derived from another source. However, all three types of clasts are remarkably similar under a microscope (Figures 9 a and 9 b). The clasts like the one shown in Figure 10 indicate that all the clasts may have been derived from one source. The clast composition is monomictic and the deposits could have resulted from a Peleean type collapse of a lava dome (see chapter 12.). The microscopic and megascopic appearance of the clasts produced may have varied for several reasons. A dome can have a microcrystalline groundmass at the interior parts where as the surface is glassy due to more rapid cooling. A surface can also be oxidized. Phenocryst contents can vary due to flow differentiation and settling. Devitrification and alternation can have a variable intensity. Flow-banding is common in viscous lavas and those can be a distinctive alternation of altered and/or devitrified layers (Cas and Wright, 1987, p. 84). The clast shown in Figure 10 is probably a piece of such a flow-banded rock.

These results suggest, however, that oxidation has a stronger influence on the petrography of the clasts in the Fe- and Mg-richer basaltic andesites (Figure 8) than in andesites; the difference in the appearance of samples IH3 BLACK or IH3 GRAY, compared to IH3 RED is small. Also, in spite of the distinctively different megascopic appearance of the devitrified/ altered sample IH3 GRAY compared to the other two samples, the difference is much less distinctive under the microscope.
Figure 9a. Photomicrograph of breccia clast sample IH3 BLACK from outcrop 3 at Indian Hollow (Appendix 2). Major phenocrysts are plagioclase and oxyhornblende. Accessory phenocrysts are clinopyroxene (augite), biotite and opaques (titanomagnetite?). Largest plagioclase phenocrysts are approximately 2 mm long. Nicols parallel.
Figure 9b. Photomicrograph of breccia clast sample IH3 GRAY from outcrop 3 at Indian Hollow (Appendix 2). Major phenocrysts are plagioclase and amphibole. Clinopyroxene (augite), biotite and opaque (titanomagnetite?) phenocrysts are present in accessory amounts. Largest plagioclase phenocrysts are approximately 2 mm long. Parallel nicols.
Figure 10. Breccia clasts from outcrop 3 at Indian Hollow (Appendix 2). Scale in centimeters.
The conglomerate pebble and cobble samples in Tables 5 and 6 illustrate another problem involved with the distinction of the source rocks. The rock types were distinguished based on a combination of textural characteristics such as phenocryst mineralogy and the size of the phenocrysts (Table 5). However, all these samples have very similar modes (Table 6).

Attempts were made to distinguish different source rocks based on the proportion of plagioclase phenocrysts having sieve-textured cores or rims, those showing complex twinning, and those having rounded corners or embayments. However, these features appear to have a gradational distribution and no clear grouping was found.

In matrix samples from coarse-grained units and volcanic sandstone samples, clasts of different andesite types were defined mainly by comparing crystallinity and other textural characteristics of groundmass. Most volcaniclastic sandstone samples contain 5-15 percent sand- or granule-sized pumice clasts. Pumiceous texture can be difficult to recognize and thin sections made with blue epoxy were most useful. Sand-sized andesite clasts are commonly poorly to moderately vesicular. Their shapes are more controlled by fracture surfaces than vesicle walls, thus giving a relatively blocky and angular geometry to both pumice and dense andesite fragments (Heiken and Wohlets, 1991).

Welding of clasts was not observed and such breccia clasts that could be fragments of previously welded rocks, while present, are uncommon.

By comparing the roundness of primary pyroclasts and fluvial pebbles 8-32 mm in diameter, Scott (1988) indicated that the debris flow deposits contained a significant portion of clasts which were introduced to the flow by basal erosion. The portion (the bulking factor of Scott, 1988) varied from 15 percent in Scott's (1988) flood-plain facies to 48 percent in his channel facies. In some of the pre-1980 deposits in the surroundings of Mt. St. Helens, the bulking factors of the channeled debris flows were as high as 70 percent. As a result of bulking and sorting, the composition of the whole clast range changed during the transportation. According to Walker (1985), coarse proximal co-ignimbrite breccias (that were coarse equivalents of Sparks layer 2a) also can contain clasts incorporated into the flow by erosion. Pyroclastic deposits commonly contain fragments of previously existing rocks, i.e. lithic fragments. Such lithic fragments may be co-magmatic to juvenile breccias and therefore, their distinction may be difficult. Data from various coarse co-ignimbrite breccias are not available.
However, Smith and Roobol (1990, p. 30) reported clast compositions from recent block and ash and scoria flow deposits and associated surge and fall deposits. The portion of lithic clasts among the sand-sized clasts is typically clearly less than 10 percent, even in lithic cobble- and block-rich beds. Commonly the sand-range contains only a few percent lithic fragments.

Because sand-sized grains are less easily rounded during fluvial transportation, the distinction of bulked-up material is less reliable from the matrix samples consisting mostly of granules and sand-sized clasts. Also a small thin section does not necessarily provide a representative sample of the over-all clast composition of the entire flow. Taking into account the results of Scott (1988) and Smith and Roobol (1990), over 10 percent of foreign clasts should indicate a polymictic composition and probable origin as a reworked deposit.

Clast compositions of some of the studied matrix samples are listed in Table 7. Taking into account variations in devitrification, oxidation, alteration, flow differentiation, cooling rate and the more or less random fragmentation that may affect the texture of volcaniclasts, the comparison of clast types from different samples is probably not reliable. However, many of the samples contain both andesite and dacite clasts and even in those samples consisting almost entirely of andesite clasts, different clast types that do not grade to one other can be distinguished. Therefore, many of the samples appear to have a polymictic composition suggesting a reworked origin.
Table 5. The proportions and textural characteristics of clast types A through H of roadcut 1 at Jordanelle (Appendix 1).

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Table 6. Modes of breccia clasts from roadcut 1.

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<td>0.1</td>
<td>0.9</td>
<td>0.1</td>
<td>0.0</td>
<td>100</td>
</tr>
<tr>
<td>JORRC1 E</td>
<td>55.6</td>
<td>24.1</td>
<td>18.5</td>
<td>0.3</td>
<td>0.0</td>
<td>0.2</td>
<td>0.2</td>
<td>1.1</td>
<td>0.0</td>
<td>100</td>
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<tr>
<td>JORRC1 F</td>
<td>55.5</td>
<td>26.7</td>
<td>14.7</td>
<td>0.8</td>
<td>0.0</td>
<td>0.0</td>
<td>1.3</td>
<td>1.0</td>
<td>0.0</td>
<td>100</td>
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<td>JORRC1 G</td>
<td>64.5</td>
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<td>7.0</td>
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<td>0.0</td>
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<td>0.5</td>
<td>3.4</td>
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<tr>
<td>JORRC1 H</td>
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<td>18.8</td>
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<td>0.0</td>
<td>2</td>
<td>2.5</td>
<td>0.0</td>
<td>100</td>
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</table>

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Table 7. Grain compositions of the breccia matrix ($\varnothing < 0.5$ cm) and volcanic sandstone samples. Minimum 600 points.

<table>
<thead>
<tr>
<th>SAMPLE</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td>JOR RC1 C (breccia matrix)</td>
<td></td>
</tr>
<tr>
<td>plagioclase-porphyritic andesite, glassy, moderately altered groundmass with small feldspar laths ($\varnothing &lt; 0.01$ mm)</td>
<td>24.9</td>
</tr>
<tr>
<td>plagioclase-hornblende-porphyritic andesite, strongly altered groundmass</td>
<td>15.3</td>
</tr>
<tr>
<td>plagioclase-porphyritic andesite, vitric groundmass with small feldspar laths</td>
<td>12.9</td>
</tr>
<tr>
<td>plagioclase</td>
<td>14.5</td>
</tr>
<tr>
<td>amphibole</td>
<td>7.4</td>
</tr>
<tr>
<td>pyroxene (including pseudomorphs consisting of opaque and micaceous mesh)</td>
<td>0.8</td>
</tr>
<tr>
<td>intergranular matrix ($\varnothing &lt; 0.06$ mm)</td>
<td>24.1</td>
</tr>
<tr>
<td>JOR RC1 1 (volcanic sandstone)</td>
<td></td>
</tr>
<tr>
<td>plagioclase-amphibole-porphyritic andesite, feldspar microlites and small plagioclase laths ($\varnothing &lt; 0.01$ mm) in glassy groundmass, porosity varies from highly vesicular to pumiceous (distinctive pumice ca. 20 % of the sample)</td>
<td>15.3</td>
</tr>
<tr>
<td>(plagioclase-amphibole-porphyritic) andesitic pumice, (pumiceos variations of the previous clast type?)</td>
<td>20.0</td>
</tr>
<tr>
<td>plagioclase-amphibole-porphyritic andesite, strongly altered variations of the clast type above?</td>
<td>10.7</td>
</tr>
<tr>
<td>plagioclase-amphibole- or amphibole-porphyritic andesite, amphibole commonly resorbed, glassy groundmass, palagonitized to brown, sporadic small plagioclase crystals</td>
<td>16.5</td>
</tr>
<tr>
<td>plagioclase porphyritic andesite, sparsely vesicular, microcrystalline groundmass ($\varnothing &lt; 0.01$ mm) with feldspar, and accessory amphibole and opaque microcrystals.</td>
<td>5.7</td>
</tr>
<tr>
<td>altered grains</td>
<td>7.5</td>
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<tr>
<td>intergranular matrix ($\varnothing &lt; 0.06$ mm)</td>
<td>10.7</td>
</tr>
<tr>
<td>plagioclase</td>
<td>7.3</td>
</tr>
<tr>
<td>amphibole</td>
<td>5.0</td>
</tr>
<tr>
<td>biotite</td>
<td>1.3</td>
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Table 7 (continued)

<table>
<thead>
<tr>
<th>Location</th>
<th>Description</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>JOR OC2 2 (breccia matrix)</td>
<td>plagioclase-amphibole ±biotite-porphyritic andesites, glassy poorly to moderately altered groundmass, moderately vesicular to highly vesicular; (plagioclase-amphibole ±biotite-porphyritic) andesitic pumice, plagioclase-amphibole ±biotite-porphyritic andesites, microcrystalline groundmass mafic volcanic fragments mudstone fragments plagioclase amphibole intergranular matrix (Ø&lt; 0,06 mm)</td>
<td>51.2</td>
</tr>
<tr>
<td>JOR RC2 1 (breccia matrix)</td>
<td>plagioclase-amphibole-biotite-porphyritic andesitic pumice plagioclase-amphibole-biotite-quartz porphyritic, sparsely vesicular dacite plagioclase-oxyhornblende-porphyritic andesite, altered microcrystalline groundmass plagioclase-hornblende-porphyritic andesite, altered microcrystalline groundmass plagioclase amphibole quartz (volcanic) intergranular matrix (Ø&lt; 0,06 mm)</td>
<td>36.5</td>
</tr>
<tr>
<td>JOR RC2 2 (breccia matrix)</td>
<td>(plagioclase-amphibole -quartz ±biotite-porphyritic) dacitic pumice (plagioclase-amphibole ±biotite-porphyritic andesite, sparsely vesicular, altered brownish groundmass plagioclase amphibole intergranular matrix (Ø&lt; 0,06 mm)</td>
<td>52.2</td>
</tr>
<tr>
<td>JOR RC3 1 (breccia matrix)</td>
<td>(plagioclase-biotite-quartz-porphyritic) dacitic pumice amphibole-plagioclase-porphyritic andesite, microcrystalline groundmass amphibole-plagioclase-porphyritic andesite, strongly altered plagioclase ±oxyhornblende porphyritic andesite, altered glassy groundmass plagioclase ±amphibole ±clinopyroxene-porphyritic andesite, moderately to strongly altered vitric groundmass sedimentary rock fragments plagioclase amphibole intergranular matrix (Ø&lt; 0,06 mm)</td>
<td>35.2</td>
</tr>
<tr>
<td>JOR RC6 1 (volcanic ss)</td>
<td>Plagioclase-amphibole-biotite-porphyritic andesite, vitric variably altered groundmass</td>
<td>27.8</td>
</tr>
<tr>
<td>------------------------</td>
<td>----------------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td></td>
<td>Plagioclase-amphibole-opaque-porphyritic andesite, moderately altered vitric groundmass</td>
<td>15.8</td>
</tr>
<tr>
<td></td>
<td>Plagioclase-oxidized-amphibole-opaque-porphyritic andesite, microcrystalline groundmass with abundant small (0 &lt; 0.01 mm) feldspar laths</td>
<td>12.5</td>
</tr>
<tr>
<td></td>
<td>Plagioclase porphyritic andesite, devitrified chert-like groundmass</td>
<td>5.5</td>
</tr>
<tr>
<td></td>
<td>Plagioclase-biotite-porphyritic andesite, vitric variably altered groundmass with abundant small (0 &lt; 0.01 mm) feldspar laths</td>
<td>4.5</td>
</tr>
<tr>
<td></td>
<td>Plagioclase-amphibole-porphyritic andesite strongly altered groundmass</td>
<td>9.0</td>
</tr>
<tr>
<td></td>
<td>Plagioclase</td>
<td>8.8</td>
</tr>
<tr>
<td></td>
<td>Amphibole</td>
<td>4.3</td>
</tr>
<tr>
<td></td>
<td>Biotite</td>
<td>0.8</td>
</tr>
<tr>
<td></td>
<td>Intergranular matrix (0 &lt; 0.06 mm)</td>
<td>10.8</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>JOR OC6 1 (volcanic ss)</th>
<th>Pumice</th>
<th>10.0</th>
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<tbody>
<tr>
<td></td>
<td>Amphibole(oxyhomblende)-plagioclase-porphyritic andesite, microcrystalline groundmass</td>
<td>13.3</td>
</tr>
<tr>
<td></td>
<td>Plagioclase ±oxyhomblende porphyritic andesite, altered groundmass</td>
<td>7.5</td>
</tr>
<tr>
<td></td>
<td>Plagioclase-homblende ±biotite ± quartz-porphyritic andesite or</td>
<td>51.2</td>
</tr>
<tr>
<td></td>
<td>Homblende-plagioclase±biotite+quartz-porphyritic andesite, altered groundmass</td>
<td>1.7</td>
</tr>
<tr>
<td></td>
<td>Plagioclase</td>
<td>1.7</td>
</tr>
<tr>
<td></td>
<td>Intergranular matrix (0 &lt; 0.06 mm)</td>
<td>16.3</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>JOR OC12 (breccia matrix)</th>
<th>Plagioclase ± clinopyroxene ± amphibole- porphyritic andesite, a few large opaques (0 1.15 mm) badly altered to non-altered vitric groundmass, sporadic mafic microlites</th>
<th>63.2</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Plagioclase ±amphibole ± clinopyroxene-porphyritic andesite, plagioclase phenocrysts Ø 1.5-2 mm, needle-shaped amphibole phenocrysts, abundant quenched plagioclase microcrystals in vitric groundmass</td>
<td>4.5</td>
</tr>
<tr>
<td></td>
<td>Plagioclase</td>
<td>11.9</td>
</tr>
<tr>
<td></td>
<td>Amphibole</td>
<td>4.8</td>
</tr>
<tr>
<td></td>
<td>Clinopyroxene</td>
<td>3.8</td>
</tr>
<tr>
<td></td>
<td>Opal</td>
<td>3.5</td>
</tr>
<tr>
<td></td>
<td>Intergranular matrix (0 &lt; 0.06 mm)</td>
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<table>
<thead>
<tr>
<th>HH 3 (breccia matrix)</th>
<th>Plagioclase-clinopyroxene-amphibole-biotite±opaque-porphyritic andesite, scattered small orthopyroxene, crystals (Ø &lt; 0.03 mm), microcrystalline groundmass</th>
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<tr>
<td></td>
<td>Strongly oxidized variations of the previous clast type</td>
<td>19.3</td>
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<tr>
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<td>Plagioclase-biotite-porphyritic andesitic pumice</td>
<td>34.1</td>
</tr>
<tr>
<td></td>
<td>Amphibole</td>
<td>5.7</td>
</tr>
<tr>
<td></td>
<td>Clinopyroxene</td>
<td>4.5</td>
</tr>
<tr>
<td></td>
<td>Biotite</td>
<td>1.1</td>
</tr>
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<td></td>
<td>Intergranular matrix (0 &lt; 0.06 mm)</td>
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Table 7 (continued)

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<thead>
<tr>
<th>Sample</th>
<th>Description</th>
<th>Data</th>
</tr>
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<tr>
<td>5 I (breccia matrix)</td>
<td>plagioclase-amphibole-clinopyroxene ± biotite porphyritic andesite, pheocrysts small (Ø 0-1-0.7 mm, vitric altered groundmass</td>
<td>36.5</td>
</tr>
<tr>
<td></td>
<td>plagioclase porphyritic andesite, microcrystalline groundmass</td>
<td>3.8</td>
</tr>
<tr>
<td></td>
<td>plagioclase-porphyritic andesite, abundant small feldspar crystals</td>
<td>7.3</td>
</tr>
<tr>
<td></td>
<td>plagioclase-amphibole-clinopyroxene-porphyritic andesite, vesicularity varies from moderately vesicular</td>
<td>14.4</td>
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<tr>
<td></td>
<td>andesitic pumice (distinctly pumiceous variations of the previous clast type)</td>
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<tr>
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<tr>
<td></td>
<td>amphibole</td>
<td>1.7</td>
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<tr>
<td></td>
<td>intergranular matrix (Ø&lt; 0.06 mm)</td>
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</tr>
<tr>
<td>7 (breccia matrix)</td>
<td>amphibole plagioclase ± clinopyroxene porphyritic andesite, altered, grayish-brown-dark brown groundmass</td>
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</tr>
<tr>
<td></td>
<td>plagioclase-amphibole-biotite ± clinopyroxene, light gray altered microcrystalline groundmass</td>
<td>19.3</td>
</tr>
<tr>
<td></td>
<td>volcanic sandstone fragments</td>
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</tr>
<tr>
<td></td>
<td>amphibole-plagioclase porphyritic andesite, glassy yellowish-brown groundmass altered grains</td>
<td>4.8</td>
</tr>
<tr>
<td></td>
<td>plagioclase-amphibole-biotite porphyritic andesite, altered groundmass</td>
<td>16.2</td>
</tr>
<tr>
<td></td>
<td>(plagioclase-amphibole-porphyritic) altered andesitic pumice</td>
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<td></td>
<td>plagioclase</td>
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<tr>
<td></td>
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<td>5.5</td>
</tr>
<tr>
<td></td>
<td>SCC 11 (breccia matrix)</td>
<td>plagioclase-amphibole-opaque ± clinopyroxene-porphyritic andesite, microcrystalline groundmass</td>
</tr>
<tr>
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<td>plagioclase-amphibole ± clinopyroxene-porphyritic andesite, vitric, partly palagonitized groundmass</td>
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<tr>
<td></td>
<td>pumice</td>
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<td>volcanic sandstone fragments</td>
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<td></td>
<td>biotite</td>
<td>1.2</td>
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<td></td>
<td>intergranular matrix (Ø&lt; 0.06 mm)</td>
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<td>SCC 11 (breccia matrix)</td>
<td>quartz sandstone fragments</td>
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<tr>
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<td>plagioclase-amphibole-porphyritic andesite, microcrystalline groundmass</td>
<td>9.0</td>
</tr>
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<td>plagioclase-amphibole-biotite-porphyritic andesite, glassy groundmass</td>
<td>5.8</td>
</tr>
<tr>
<td></td>
<td>amphibole-plagioclase porphyritic andesites, altered groundmass</td>
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<td>red sandstone fragments</td>
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<tr>
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<td>limestone (micrite) fragments</td>
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<tr>
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<td>limestone (sparite) fragments</td>
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</tr>
<tr>
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<td>plagioclase</td>
<td>12.3</td>
</tr>
<tr>
<td></td>
<td>intergranular matrix (Ø&lt; 0.06 mm) (calcareous)</td>
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Table 7 (continued)

<table>
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<tr>
<th>SCC 8 1 (breccia matrix)</th>
<th>plagioclase-amphibole-biotite-quartz porphyritic dacite</th>
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<tr>
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<td>pumice</td>
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<td></td>
<td>plagioclase-amphibole-porphyritic andesite, altered</td>
<td>3.3</td>
</tr>
<tr>
<td></td>
<td>plagioclase</td>
<td>11.6</td>
</tr>
<tr>
<td></td>
<td>quartz</td>
<td>3.1</td>
</tr>
<tr>
<td></td>
<td>strongly altered grains</td>
<td>7.3</td>
</tr>
<tr>
<td></td>
<td>intergranular matrix (Ø&lt;0.06 mm)</td>
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</table>

<table>
<thead>
<tr>
<th>SCC 15 1 (breccia amphibole-plagioclase ±biotite-porphyritic andesite and plagioclase-amphibole-matrix)</th>
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</tr>
</thead>
<tbody>
<tr>
<td>plagioclase ±biotite-porphyritic andesite, microcrystalline gray groundmass</td>
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</tr>
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<td>(plagioclase-amphibole ±biotite-porphyritic) andesitic pumice</td>
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</tr>
<tr>
<td>plagioclase</td>
<td>0.1</td>
</tr>
<tr>
<td>amphibole</td>
<td>20.8</td>
</tr>
</tbody>
</table>
10. Deposits at Jordanelle

The studied outcrops and roadcuts at Jordanelle are located in the Heber quadrangle in sections 31-33, T.2S. and R.5.E, and sections 4-7 and 18, T.3S. and R.5E. The roadcuts are along new highway 189. A map of the Jordanelle area is included as Appendix 1.

10.1 Roadcut 1

At what is apparently the lowest exposure of the Jordanelle profile (Figure 11, roadcut 1) deposits are thickly or very thickly bedded polymictic volcanic conglomerates and breccias with shallow channels and intercalated thin to medium-thin volcanic sandstone beds. At the northern end of the roadcut is a subvertical fracture. On the north side of this fracture, the deposits are altered; even large cobbles are soft and break easily. Some gently and moderately dipping fractures are also found at the southern end of the roadcut. Alteration along fractures has produced yellow phyllosilicates in a few centimeters wide slightly lighter-colored zones which have gradual contacts with the unaltered matrix. Some of the reactive solutions have flowed along bedding planes.

10.1.1 Clast lithologies

Gravel-sized clasts are mainly plagioclase-amphibole porphyritic andesites. Accessory clasts include small granules of light gray calcareous sedimentary rocks. Volcanic rock clasts are dense or are vesicle-poor. Some of them are partially altered to yellowish phyllosilicate. Such partially altered clasts are sharply surrounded by an unaltered matrix. Therefore, the alteration appears to have taken place prior to deposition. Altered clasts are unlikely to remain intact if transported by traction or by sedimentary mass flows in which the transportation competence is derived from turbulence, dispersive stress or friction between the clasts. However, altered clasts can be transported, for example, by debris flows in which clast interactions are less intense and the clast support is due to cohesion or buoyancy. It should be noted also that here the clasts are not thoroughly altered, and thus have some inherent strength.

The proportions and textural characteristics of different clast types present at roadcut 1 are shown in Table 5. The proportions were estimated by attaching a tape
measure, 1 m long, on three arbitrarily chosen lines in a clast-supported conglomerate/breccia bed and by counting the proportions of different clast types in the traverses. Clasts smaller than 0.5 centimeter in diameter were counted as a part of the matrix.

Sandstone beds have heterogeneous clast compositions similar to the sandy matrix of the conglomerates. Thin section studies show that the sand-sized clasts are predominantly dense or sparsely vesicular andesites, but pumice clasts are present in amounts as high as 20 percent (Figure 12, Samples JOR RC1 C and JOR RC 1 in Table 7). Inhomogeneities in the muddy matrix suggest that it is mostly diagenetic (Dickinson, 1970).
Figure 11. Volcanic conglomerates/breccias and sandstones in roadcut 1 at Jordanelle (Appendix 1). Deposits are slightly altered along joints, minor faults and bedding (bottom right). Clasts are polymictic. Some of the clasts are altered prior to deposition (bottom left). Notebook is 30 cm long.
Figure 12. Photomicrograph of a volcaniclastic sandstone at Jordanelle, roadcut 1 (Appendix 1). Examples of various clast types: blue tinted pumiceous to highly vesicular (top, left; center-left) and moderately vesicular plagioclase-amphibole-porphyritic clasts (center, top); amphibole-plagioclase-porphyritic clasts with brown palagonitized groundmass (top, center; right, bottom); plagioclase-porphyritic clasts with microcrystalline groundmass (top, right corner). Clast composition is listed in Table 7, sample JOR RC 1.
10.1.2 Descriptions

Matrix-supported conglomerates (breccias?)

Matrix-supported conglomerates occur as thick beds in which the lower boundaries are sharp. Subangular and subrounded pebbles, cobbles and boulders are supported in a sandy matrix containing little mud. Cobbles and boulders form clusters in which several clasts are in contact.

Inversely graded, matrix-supported conglomerates (breccias?)

Thick inversely graded, matrix-supported conglomerates have thin basal zones lacking cobbles. Some of the larger clasts have point contacts. The inverse grading takes place as a relatively rapid jump in the proportion of larger clasts above a thin basal layer. The clasts are polymictic and poorly rounded. The sandy matrix contains abundant granules and minor mud.

Clast-supported conglomerates (breccias?)

Clast-supported conglomerates have sharp or gently channeled bases. Thick to very thick beds are massive or crudely bedded with slight variations in grain size. The beds may include a poorly developed, laterally discontinuous and thin inversely graded basal zone. Subangular or subrounded clasts are typically 5 to 15 cm in diameter, but also present are large cobbles and boulders (maximum Ø 70 cm). The matrix is sandy with abundant granules and small pebbles. Less than 5 percent of the sample is muddy matrix.

Normally graded sandstones

A few normally graded, coarse to medium-coarse sandstones are present as medium-thick beds in discontinuous sand lenses between channeled conglomerates. Grading can be well-developed through the entire thickness of the bed or limited to the upper third. In one of the graded beds, the upper part of the bed contains few (1-3 percent) strongly altered grayish-white pebbles dispersed in a medium-coarse sand; the lower part of the bed is fine gravel or coarse sand. The granules in this coarse proportion are similar to the volcanic rocks found in the deposits as larger clasts. Grayish-white, strongly altered pebbles were probably pumice clasts that had similar settling velocities to smaller but denser sand-sized andesite and crystal clasts.
Horizontally stratified sandstones

Very thin or thin horizontally stratified sandstone layers are commonly eroded by coarser beds. The thickness of internal stratification varies from lamination to very thin bedding. Stratification is distinctive but not sharp, being defined by slight variations in grain size. The grain size is generally coarse or medium sand. The beds contain a few larger granules at their bases.

10.1.3 Interpretation of roadcut 1

The cobble and pebble samples have a very similar mineralogy but the observed variations in their textures indicate a polymictic composition. Both the matrices of the conglomerates/breccias and sandstones contain abundant pumice. However, pumice is scarce as pebbles or larger clasts.

The polymictic composition of the clasts and the varying degree of predepositional alteration indicates a reworked origin. In conglomerate beds, various clast types are uniformly mixed in a homogeneous matrix through the entire vertical thickness of the bed. This indicates that the flows have been turbulent at some point and have been bulked up by rounded fluviatile sediments. Although some non-volcanic debris flows may have been turbulent (Nemec and Steel, 1984), most likely bulking has occurred during a hyperconcentrated flow stage preceding plastic behavior of a debris flow similar to the river system in the vicinity of Mount St. Helens (Scott, 1988).

Matrix-supported massive conglomerates/breccias were most likely deposited from debris flows by cohesive freezing (Nemec and Steel, 1984; Smith and Lowe, 1991). Clast-supported clusters probably represent incipient settling of larger clasts. Inversely graded matrix-supported conglomerates were deposited by debris flows in which basal shearing was significant; in addition to having better developed inversely graded basal zones, the beds deviate from matrix-supported conglomerates by having slightly more granular matrix. These characteristics suggest that clast interactions were favored at the expense of cohesion as a support mechanism (Nemec and Steel, 1984; Shultz, 1984; Smith and Lowe, 1991).

Clast-supported conglomerates have a relatively coarse matrix that could not have filtered into interclast spaces after the clast-by-clast deposition of larger pebbles and cobbles (Smith, 1986). A dilute stream flow origin is neither favored by the lack of imbrication in the conglomerates and breccias nor by the lack of traction sedimentation.
structures at the roadcut. This lithology was formed by the rapid deposition of sediments representing a range of grain sizes (Smith, 1986). Hyperconcentrated flow gravels are typically normally graded. Therefore, the clast-supported conglomerates were most likely deposited by relatively dilute debris flows that were capable of eroding underlying fine sediments as well as gravelly deposits.

Well-developed normal grading in sandy turbidity current deposits is attributed to rapid and collective deposition from suspension (Lowe, 1982). This mechanism is also called upon to explain deposition of various normally graded and graded-stratified hyperconcentrated flow deposits having great similarities to turbidites (Cas and Wright, 1987; Smith and Lowe, 1991). Similarly, the normally graded sandstones have a relatively narrow grain-size distribution and their deposition is most conveniently explained by mechanisms proposed by Lowe (1982) for subaqueous turbidity currents. Horizontally stratified sandstones lack distinctive layers with different grain sizes. Although the origin of such stratification is not fully understood, it is often recognized within "shallow braided" or "flash flood" deposits which are commonly described as hyperconcentrated deposits.

10.2 Outcrop 6

At outcrop number 6 (Appendix 1) a 20 meter thick sequence of volcanic breccias and sandstones is exposed. The volcanic rock fragments are plagioclase-hornblende porphyritic andesite with accessory rounded quartz crystals showing an undulose extinction (Figure 13). The breccias contain accessory amounts of light gray quartz-sandstone clasts.

Grayish-white clasts are strongly altered to clay which makes sampling difficult. A sample from a less altered light gray clast contains abundant vesicles (ca. 25 percent) but is not porous enough to be described as pumice. This sample has a mineralogical composition very similar to dense andesite cobbles in the same outcrop. Pumice is present in a thin section from a volcanic sandstone bed (ca. 10 percent, sample JOR OC6 1 in Table 7, Figure 14) and in a scanning electron microscope sample of tuffaceous matrix. In volcanic rock fragments, glass is partly altered to a chlorite-smectite mesh. Cobble- and boulder-sized grayish-white clasts were probably not pumice, and their alteration may be predepositional. Coarse breccia clasts are typically either poorly or highly vesicular; moderately vesicular clasts are scarce. One of the
andesite clasts has a bread-crusted surface (Figure 15). The bread-crusted clast is not bomb-shaped but could be a piece of a broken bomb. Most of the sparsely vesicular andesites, as well as the quartz sandstone fragments, are derived from previously existing rocks. The uniform mineralogical composition suggests that such clasts are contiguous or co-magmatic with grayish-white altered clasts and probably also with tuffaceous components in the matrices of breccias and associated volcanic sandstones.
Figure 13. Photomicrograph of a vesicle-poor andesite clast from outcrop 6 at Jordanelle (sample JOROC 6 2A, Table 4) with plagioclase and amphibole (hornblende) phenocrysts and accessory subhedral to euhedral opaques (titanomagnetite?), subhedral biotite and rounded quartz crystals which show undulose extinction. Since the sample is lacking any evidence of brittle or ductile deformation, the quartz crystals are more likely xenocrysts than strained phenocrysts. The largest plagioclase phenocrysts are 3 mm in diameter.
Figure 14. Photomicrograph of a volcanic sandstone (unit 6, outcrop 6 at Jordanelle, Appendix 1) with scattered pumice clasts (partly filled by blue tinted epoxy). Majority of the clasts are dense sparsely vesicular clasts and consist of amphibole-plagioclase-porphyritic clasts with microcrystalline groundmass (center, left), plagioclase±oxyhornblende-porphyritic clasts with altered groundmass (top, right) and plagioclase-hornblende- or hornblende-plagioclase-porphyritic (±biotite±quartz) clast with altered groundmass (center; bottom, left). Largest clasts are ca. 3 mm in diameter.
10.2.1 Unit descriptions

Unit 1 is an inversely to normally graded volcanic breccia with a minimum thickness of 4.2 m. The lowest 0.8 m has up to 15 cm long (median $\varnothing$ 7 cm) grayish-white altered clasts supported by a tuffaceous matrix. Sparsely vesicular andesite clasts are less than 3 cm long (median $\varnothing$ 1.5 cm). The rock grades upward to a 2.5 m thick coarser middle part in which the texture becomes clast-supported. Clasts include both altered light gray clasts and sparsely vesicular andesite. In the upper part, the proportion of larger clasts (pebbles and cobbles) decreases and in the uppermost 0.4 m they are absent. The unit contains accessory amounts of light gray quartz sandstone fragments.

Units 2-10 are shown in Figures 16 a and 16 b. Unit 2 is a 2 m thick volcanic breccia. Clasts are up to 25 cm long with median size of 10 cm; and they vary from vesicle-poor to vesicle-rich. The matrix contains granules and small pebbles which make the texture clast-supported. The unit contains accessory amounts of quartz sandstone fragments.

Unit 3 is a 1.8 m thick matrix-supported volcanic breccia that has been eroded by the overlying unit. Its lower boundary is sharp and dips gently to the east. The largest sparsely vesicular andesite clasts are 5 cm long, but the median size is ca. 2 cm. The maximum diameter of altered, grayish-white clasts is 35 cm and the median size is approximately 10 cm. Sparsely vesicular andesite clasts are somewhat concentrated towards the base but the top of the bed is more clearly normally graded. The uppermost 20 cm is tuffaceous. Granule- or small pebble-sized dense andesite clasts and larger grayish-white altered clasts are scattered in the unit.

Unit 4 is a thin, medium-coarse to coarse-grained crystal-rich volcanic sandstone bed. According to megascopic observations, the clasts are mainly plagioclase with some amphibole.

Unit 5 is a few centimeter thick, fine volcanic sandstone with a sharp lower contact. Unit 6 is a 0.5 cm thick volcanic sandstone containing a few dispersed pumiceous granules and small pebbles. The lower contact is sharp. The bed is discontinuous due to erosion by the overlying bed. A sample of unit 6 (sample JOR OC6 1 in Table 7, Figure
14) includes a few pumice clasts (10 percent) and volcanic rock fragments with microcrystalline groundmass (10-15 percent).

Unit 7 is an 0.7 m thick altered grayish-white clast-rich volcanic breccia with an erosional base. The largest altered grayish-white clasts are 55 cm in diameter. The bed is relatively clast-rich and the largest clasts are commonly in contact with each other. However, most of the pebbles and cobbles are supported by matrix. The matrix is relatively granular, with abundant volcanic rock and crystal fragments.
Figure 15. An andesite clast with bread-crust surface in a volcanic breccia (unit 2 from outcrop 6 at Jordanelle).
Figure 16a. Deposits from unit 2 to unit 10 of outcrop 6. Scale bar is 1.55 m long.
Unit 8 is a 2.5 m thick breccia with poorly developed inverse to normal grading. Boundaries are sharp. Coarse clasts form approximately 70 percent of the bed. The texture is matrix-supported with very coarse matrix rather than clast-supported. In the 1 m thick lower part, granule- and small pebble-sized dense to sparsely vesicular andesite clasts show poorly developed inverse grading although it contains some scattered large altered blocks. In the central part, the size of the largest andesite clasts reaches 15 cm; the median size is 3-4 cm. In the upper part, the proportion of larger clasts decreases and the uppermost 0.2 m does not contain dense andesite fragments. The largest altered clasts are 7 cm long.

Units 9 and 10 are similar to units 5 and 6. Unit 10, over 7 m thick, is coarse matrix-supported breccia in which the largest dense to sparsely vesicular andesite blocks are 1 m in diameter. An approximately 1 m thick basal zone is inversely graded, with blocks up to 40 cm long. The matrix is relatively granular with abundant crystal fragments and volcanic rock fragments.
10.2.2 Interpretation of outcrop 6

The breccias do not show well-developed layer 2a or 2b structures of Sparks (1976). In spite of being relatively monomictic, the breccias are likely debris flow deposits. Some of the vesicular andesite blocks are strongly altered or weathered. Since some of the sand- and granule-sized pumice clasts, with a relatively larger effective surface area are relatively intact, the alteration of larger clasts probably occurred in the source area. The poorly developed inverse grading resulted from basal shear pressure. The granular sand matrix suggests that plug behavior of the massive non-graded part was induced by intergranular friction. Some of the beds with normal grading of sparsely vesicular andesite clasts seem to have a slightly increased content of grayish-white highly vesicular clasts in their upper parts. This vaguely developed grading could be similar to density grading in pyroclastic flows in the layer 2b of Sparks (1976). Alternatively and perhaps more likely, this pattern developed in the more fluidal upper parts where denser clasts started to settle and maintained a higher clast content in lower parts where less durable clasts are more effectively abraded due to more intense intergranular shear. Less durable, previously altered clasts are better preserved in the upper part.

Volcaniclastic sandstones are also difficult to interpret. Unit 4 is a thin crystal-rich bed and could have resulted from gravity settling of crystals from an ash cloud. Pyroclastic flows are often associated with ash clouds that deposit fine surges or fall deposits. As emphasized by Cas and Wright (1987, p. 126), "one would intuitively expect such deposits to be enriched in vitric particles". However, some deposits are crystal-rich and "this must be due to further gravity segregation within the ash cloud". In spite of the fact that thin fine volcaniclastic mudstone layers (units 5 and 9) are found below thick massive volcanic pebbly sandstones (units 6 and 10), such deposits do not seem to occur together higher in the stratigraphy. Therefore these deposits are probably unrelated. If pyroclastic, these fine and relatively well sorted layers probably represent ash cloud fall or distal surge deposits. Such layers could be low-energy fluvial or even lake deposits. However, such interpretation is not supported by the lack of tractional sedimentary structures in the overall sequence (Nemec and Steel, 1984; Smith and Lowe, 1990).

A sample of unit 6 includes a few pumice clasts (10 percent) and volcanic rock fragments with microcrystalline groundmass (10-15 percent). Sedimentary rock
fragments are found in accessory amounts. The clast composition suggests a reworked origin for the bed. A pyroclastic origin cannot be ruled out but it is not favored because of the lack of grading patterns typical of flows and by the lack of stratification typical of pyroclastic falls (Sparks et al., 1976; Lajoie, 1984; Fisher and Schmincke, 1984). Unit 6 seems to be too thick to be a sole layer of the overlying debris flow. Although nongraded, it may be comparable to Scott's (1988) pebble-carrying peak flows that covered coarse basal debris flow deposits and spilled over to floodplain areas.

10.3 Outcrop 1

10.3.1 General description of outcrop 1

Outcrops 1a and 1b (Appendix 1) stratigraphically overlie outcrop 6. The covered interval between outcrop 6 and outcrop 1a is approximately 30 m. At these outcrops the deposits have relatively monomictic clast compositions. Sparsely vesicular clasts are mainly plagioclase porphyritic andesite containing some amphibole phenocrysts. At outcrop 1a the rocks are altered grayish-white clast-rich deposits also containing sparsely vesicular andesite fragments (median $\bar{D}$ 3 cm, maximum $\bar{D}$ 10 cm). The deposits seem to be thickly bedded but intense weathering obscures detailed observations.

Overlying a covered interval of 15-20 m, at outcrop 1b, (Appendix 1), is a 3 m thick sequence of thickly bedded breccias and thin- to medium-bedded volcanic sandstones (Figure 17).
Figure 17. Stratigraphic column for outcrop 1b (Appendix 1).
10.3.2 Breccias

A few beds have inversely graded basal divisions which are followed abruptly by partly clast-supported cobbly zones. In the upper parts of the beds, the amounts of andesite clasts decrease as the amounts of grayish-white altered clasts increase.

Excluding the uppermost, partly exposed bed, andesite clasts are characteristically plagioclase porphyritic; the color of the groundmass varies from reddish-brown to dark gray. One breccia unit (Unit 14, Figure 17) contains a thin inversely graded basal division followed by a nongraded major part with both grayish-white altered clasts and sparsely vesicular andesite clasts.

In spite of the observed crude grading of lower density components towards the top of unit 4, the breccia beds were probably not deposited by expanded (partially fluidized) pyroclastic flows or non-expanded, poorly fluidized flows (Wilson, 1980). Based on comparisons with altered clasts in lower deposits, the whitish altered clasts are probably not pumice. Their crude concentration in upper parts of the beds may be due to incipient density grading similar to pyroclastic flows (Sparks, 1976) and fluidal debris flows (Scott, 1988) or it may result from intense shearing and cataclasis of softer clasts near the base. The large amount of sparsely vesicular andesite clasts is similar to block-and-ash-flow deposits. However, the deposits contain abundant large lithics (i.e. pieces of previously existing rocks) and are lacking gas escape structures and hot in situ brecciated clasts. One of the breccia layers has a pinkish-gray matrix (unit 15). Since the above-mentioned characteristics of high temperatures during deposition are lacking, the bed may have been deposited by a debris flow. Debris flows may have even more likely deposited the partially exposed unit 16 which has an inversely graded base, a coarse sandy matrix and a polymictic clast composition.

10.3.3 Graded volcanic sandstones

Beds 1, 2 and 10 (Figure 17) are graded. They have either a thin or missing inversely graded basal zone in which grading involves both dense and grayish-white altered fragments. The major parts of the beds display coarse-tail grading with normal grading of dense clasts and inverse grading of low density clasts.

Many of these low density clasts probably were not pumice, grading patterns are not very distinctive and beds lack gas escape structures. Therefore, deposition cannot be definitely attributed to partially fluidized pyroclastic flows (Sparks et al., 1986). Scott
(1988) has recorded local concentration of wood, pumice and other low density clasts on or near bed surfaces in his otherwise inversely to normally graded floodplain deposits. Likewise, deposition may have taken place from fluidal debris flows.

10.3.4 Volcanic pebbly sandstones

Strongly altered, originally highly vesicular clasts form lag-like layers in unit 13 that are as thin as one clast thick but are also found as a 7 cm thick structureless bed (unit 6).

These beds may be thin pyroclastic deposits (fall or flow), but they may have also resulted from reworking processes.

10.3.5 Volcanic sandstones

A thin tuffaceous bed contains dispersed angular to subangular lithic fragments and subangular to subrounded pumiceous clasts. The bed lacks internal stratification, which suggests a pyroclastic flow origin (Lajoie, 1984).

A medium-thick medium-coarse tuff bed (unit 8) has a 10 cm thick massive basal division that grades upward to slightly finer tuff in which pumiceous clasts are concentrated in stringers and lenses several centimeters to over a decimeter long. The unit is moderately to poorly sorted. Another crudely stratified unit (12, Figure 17) is a fine tuff with coarser tuff stringers a few centimeters long. Similar units were found higher in the stratigraphy and their depositional mechanisms will be discussed later.

Slightly better sorting of unit 11 suggests that this horizontally stratified unit could be a fall deposit.

10.4 Outcrops 3 and 5

Roughly 10 m stratigraphically above outcrop 1 are outcrops 5 and 3 (Appendix 1). These deposits are thick breccias with a few thin interbeds of coarse volcanic sandstone. The breccias are clast-supported and a few of them have thin inversely graded basal zones. The matrix is granular sand. The lower boundaries are sharp and/or represent erosion surfaces. The average grain size is in the small pebble size, and larger clasts have diameters as large as 10 cm. The clasts are reddish-gray, fine-grained plagioclase-porphyritic andesites and gray, plagioclase-amphibole porphyritic andesites.

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and altered grayish-white clasts. Some of the clasts have thin predepositional alteration or weathering rims, indicating a reworked origin. Volcanic sandstone beds are massive; some approach fine gravels. Coarse sandstone beds have polymictic sources similar to their coarse counterparts.

The breccia beds are likely deposited from relatively fluidal debris flows (Nemec and Steel, 1984; Smith, 1986; Cas and Wright, 1987). The sandy interbeds probably represent hyperconcentrated flows of volcaniclastic detritus.

10.5 Outcrop 2

10.5.1 Breccias

Thick volcanic breccias and interbedded sandstones continue higher in the local stratigraphy where several closely spaced outcrops constitute a 25 m thick sequence that consists of gentle to deep channel fills which complicate lateral correlations (outcrop 2, Appendix 1).

The clast composition is variable from polymictic to relatively monomictic beds consisting entirely of plagioclase-amphibole porphyritic andesite and grayish-white strongly altered clasts. Among the polymictic breccias, two lithologies can be outlined (Figure 18).

Inversely graded matrix-supported breccias (conglomerates?) have sharp and/or erosional bases. The clasts vary from angular to subrounded, with the largest 2.2 m long. Even such large clasts are clearly suspended above the base. In some of the beds, the cobble and boulder content increases so much that in the upper 40 percent of the bed the large clasts have point contacts. The matrix is sand-rich but also contains some mud.

Clast-supported conglomerates occur as lenses as thick as 0.5 m in which the clasts are subangular to subrounded. The sandy matrix contains abundant granules. A finer grained (pebbly) clast-supported conglomerate with a few oversized, rounded cobbles occurs as a lag at the base of a channel. The lag is overlain by stratified (low-angle) cross-bedded pebbly sandstone. The contact between the two lithologies is not clearly gradual. The deposition of these two lithologies resulted from a decrease in flow velocity that led to deposition of a fining-upward sequence (Cant and Walker, 1976).
Some matrix-supported breccia and conglomerate beds contain strongly altered grayish-white clasts, dense reddish-gray plagioclase-amphibole porphyritic andesite clasts, and abundant highly vesicular (scoria) clasts that vary in color from rusty brown to gray. However, the matrix of the same bed (sample JOR OC2 2, Table 7) appears to have been derived from a single source, for it consists entirely of angular plagioclase-amphibole-biotite porphyritic andesite fragments. Also present are a mafic plutonic rock fragment which is a piece of a cognate inclusion or a plutonic wallrock, and a mudstone fragment. Most of the clasts are plagioclase porphyritic andesite that contain little amphibole (c°z 11°) and accessory biotite phenocrysts. A small proportion of andesite clasts have a microcrystalline groundmass, but in the majority the groundmass is glassy. Many of the glassy fragments are pumice but some have hardly any vesicles (as indicated by thin sections with stained epoxy). Approximately 25-30 percent of the clasts are pumice. Volcanic rock fragments seem to be angular and blocky. Cuspate or Y-shaped shards are missing. Crystal grains are plagioclase, amphibole and biotite having compositions similar to the phenocrysts of the volcanic rock fragments. The sample contains 10-15 percent matrix that is probably mostly diagenetic smectite and zeolite mesh.

Monomictic breccias are thick to very thick; the beds tend to be crudely inversely graded (Figure 19). Grading occurs as a distinctive jump in grain size, with a pebbly matrix-supported basal part overlain by a layer of large cobbles and boulders (maximum Ø 1 m). Large clasts are commonly in point contact and protrude into overlying deposits or are overlain by a thin layer of strongly altered pebble-sized clasts.

Conglomerate beds consisting of altered light gray to grayish-white clasts are found as thin (a few centimeters to few decimeters thick) clast-supported beds or discontinuous lags and thin lenses in stratified pebbly sandstones (Figures 18-20). Such beds can show inverse to normal or normal (coarse tail) grading.
Figure 18. Polymictic deposits at outcrop 2 (Appendix 1). The partly exposed bed at the bottom is clast-supported. It is overlain by a thin clast-supported pebbly conglomerate lag which is followed by low-angle trough cross-bedded pebbly volcanic sandstone. The sandstone includes scattered oversized cobbles and is overlain by a pinkish-gray volcanic siltstone lamina. The thick uppermost bed is inversely graded and matrix-supported and contains abundant pumice as breccia clasts in the sandy matrix. Other clasts are dense andesites and andesitic scoria. The dark vertical bands are due to algae.
Figure 19. Crudely inversely graded breccia at outcrop 2 (Appendix 1). Sparsely vesicular andesite clasts are essentially derived from one source. In addition to andesite clasts, the bed contains altered light gray granules and small pebbles many of which are pumice. The breccia is overlain by a stratified volcanic sandstone containing pebble lags and low-angle cross-bedding. Note hammer for scale (top left).
Figure 20. Grading from altered clast-rich lag to crudely stratified, low-angle cross-beded pebbly sandstone (outcrop 2, Appendix 1). Note the scattered andesite pebbles. Altered light gray clasts have probably had settling velocities close to those of the sparsely vesicular andesite fragments and therefore the altered whitish clasts are probably not juvenile pumice. The coin for scale is 2.5 cm in diameter (top right).
10.5.2 Volcanic sandstones

Stratified pebbly volcanic sandstones are found as thin to thick interbeds in both polymictic and monomictic coarse deposits. These deposits contain small strongly altered clasts and sparsely vesicular andesite clasts. Granule- and pebble-sized clasts are angular to subrounded; highly vesicular clasts are more rounded than the dense clasts.

An example of a low-angle cross-bedded volcanic pebbly sandstone is shown in Figure 18. Thin clast-supported gravel lag is overlain by low-angle cross-bedded pebbly sandstone.

Some stratified volcanic sandstone beds can be traced several tens of meters. However, individual layers in these beds are less continuous and channeled. Internal stratification is planar, lenticular or forms low-angle cross-bedding. Stratification is commonly diffuse and includes altered clast-enriched layers and lenses (Figure 21). The debris has been sorted into compositionally variable layers.

Stratification is more distinctive in fine and medium-coarse volcanic sandstone channels (Figure 22). In these deposits, stratification does not mantle channel walls but is instead clearly truncated.

One bed a few decimeters thick shows inverse grading of both poorly and highly vesicular clasts through almost the entire thickness of the bed.
Figure 21. Stratification in pebbly volcanic sandstones (outcrop 2, Appendix 1). Note 2.5 cm coin for scale (bottom left).
Figure 22. Stratification in fine to medium-coarse volcanic sandstone (outcrop 2, Appendix 1). Note 2.5 cm coin for scale (bottom left)
10.5.3 Interpretation of outcrop 2

Deposits with clearly polymictic clast compositions likely resulted from reworking processes. Therefore, most of the inversely graded matrix-supported breccias can be attributed to debris flows, whereas the channeled clast-supported conglomerates can be attributed to fluidal (cohesionless) debris flows or stream flows.

The sample from matrix-supported breccia described above with polymictic clast composition but apparently homogeneous matrix contains vesicular rock fragments which were probably derived from eruptions that produced abundant pumice as well as less vesicular fragments. However, variations in the clast composition of both sand and coarser fractions suggests a reworked origin as debris flows that were able to bulk up with dense andesite pebbles and cobbles. The origin of the relatively abundant muddy matrix is problematic; it may indicate a previously weathered source, it may have been incorporated into the flow during its movement by bulking, or it may have resulted from the intense diagenetic alteration of permeable debris.

The depositional mechanism of crudely inversely graded breccias with relatively monomictic clast compositions is difficult to ascertain. Erosional bases and poor vesicularity exclude a fall origin for these deposits. They do not contain dense bombs, are not laterally continuous, and do not seem to form internally stratified cycles. None of the breccia layers show welding and they are not associated with ignimbrites. These units are, therefore, likely not to be lag breccias. The relatively monomictic composition of the dense clasts suggests a pyroclastic flow origin. The flows probably resulted from explosive dome collapses that produced dense blocks, pumice clasts and ash. Poorly developed concentrations of low density clasts could indicate poor fluidization of such flows. However, a debris flow origin cannot be completely excluded. Probably the best analogs are graded-stratified deposits described by Smith (1984) which record change from hyperconcentrated flow conditions and suspension settling to dilute stream flow and tractional sedimentation.

The alteration of light gray clasts was intense and their sampling was unsuccessful. Therefore, it is unclear if these clasts were pumice, as megascopically observed relict banding and the abundance of pumice in the matrix of associated breccias suggests. The altered light gray clast-rich normally graded beds and lags which are associated with
low-angle cross-bedded pebbly sandstone contain scattered andesite clasts. The grain sizes of these apparently dense clasts are not significantly different from those of altered clasts. This suggests that the latter did not include much pumice.

The association of horizontally stratified and low-angle cross-bedded volcanic pebbly sandstones with underlying gravel lags shown in Figures 18 and 19 reflects a slight decrease in flow velocity of a stream rather than grading in a mass flow losing its competence. Therefore, these two lithologies form a fining-upward sequence (comparable to those of Cant and Walker, 1976) formed under waning flow conditions of a possibly lahar-related stream.

The truncated stratification of fine or medium-coarse sandstones at channel walls argues against a pyroclastic fall or surge origin. Since pyroclastic deposits are massive or diffusely stratified (Lajoie, 1984; Cas and Wright, 1987, p. 484-485) the deposition of these distinctively stratified sandstones was most likely by streams. Beds showing inverse grading through their entire thickness have been described from modern deposits of Mount St. Helens (Scott, 1988). Such deposits represent transitional conditions between debris flows and hyperconcentrated flows.

10.6 Roadcut 2

10.6.1 General description

An excellent exposure of deposits younger than those already described is provided by a 300 m long roadcut that contains a 30 m thickness of variable lithologies (Appendix 1). Bedding is horizontal. At the northern end of this roadcut stratified deposits can be traced laterally for approximately 100 m. The exposure contains some deep channels and is cut by faults and joints. The faults have apparent vertical displacements of a few meters. The deposits at the southern end of roadcut 2 may represent the same stratigraphic horizon as outcrop 2, separated by 0.4 km.

10.6.2 Deposits at the southern end

At the southern end of the roadcut, thick matrix-supported polymictic breccias and conglomerates are exposed (Figure 23).

Deposition has occurred from flows which were able to carry large blocks close to their tops. The largest block is over 1.5 m in diameter. A sample from unit 2 contains
abundant pumice pebbles and vesicle-poor or dense volcanic rock fragments. Pumice is commonly formed in explosive eruptions of acid magmas. Therefore, one might expect that the pumice clasts had a more siliceous composition than the sparsely vesicular clasts shown in Figure 24. However, the pumice clasts that constitute 40 percent of the sample (JOR RC2 1, Table 7) have andesitic mineralogy whereas the most of the dense clasts (ca. 30 percent of the sample) have quartz phenocrysts (and thus, are dacite fragments). Vesicle-poor plagioclase-amphibole porphyritic andesites (ca. 10 percent) have also contributed to the distinctively polymictic clast composition.

One of the flows (unit 3, Figure 23) has eroded underlying cobbly deposits and has crude stratification close to its base. Two coarse beds are separated by a thin and discontinuous pinkish volcanic silt layer. The silt layer is distinctively finer and better sorted than the matrix of the overlying bed and therefore, it is not likely a sole layer (Scott, 1988). The overlying bed has an inversely graded basal zone indicating some shear at the base although the flow was not very erosive. This unit (Unit 3, Figure 23) is megascopically very similar to flows described earlier (The uppermost bed in Figure 18) and may represent the same stratigraphic horizon separated by 0.4 km.

The strongly erosional unit (unit 5, Figure 23) had a higher flow competence than unit 3. Similar crude stratification resulting from basal shearing has been described in non-volcanic debris flows as "pseudolamination" by Nemec and Steel (1984).

Figure 23. Stratigraphy at the southern end of roadcut 2 at Jordanelle (Appendix 1). All breccia/conglomerate units are matrix-supported. Largest boulder is more than 1.5 m in diameter. The total thickness of the represented sequence is ca. 3 m.
Figure 24. Photomicrograph a pebbly conglomerate sample (JOR RC2 1, Table 7) from the southern end of roadcut 2 at Jordanelle (Appendix 1). Parallel nicols. The largest phenocryst (plagioclase, bottom right) is 2 mm long.
10.6.3 Poorly coherent conglomerates

The deposits described above are cut by a channel that is filled with clast-supported, polymictic pebbly conglomerate and massive conglomerate (Figure 25). The deposits contain abundant older sedimentary rock clasts (chert, limestone, light-colored quartz sandstone, red sandstone) as well as buff colored volcanic tuff that is similar to the tuff of Peoa.

A partially exposed, stratified wedge consists of at least three clast-supported conglomerate beds. The beds are clearly truncated by the steep channel wall. The well-rounded and sorted texture suggests a fluviatile origin.

The massive conglomerate consists of large gray blocks of volcaniclastic-dominated conglomerate chaotically suspended in reddish sedimentary-clast rich-conglomerate. Some of the gray blocks are oxidized reddish along fractures. In spite of this evidence of postdepositional oxidation, the unit consists of two different lithologies rather than representing a homogeneous conglomerate that is incompletely oxidized. The reddish conglomerate is richer in red sandstone (Nugget Sandstone) and light-colored limestone fragments. The red conglomerate is soft, whereas the gray conglomerate blocks are more coherent. Some of the gray blocks were clearly altered to yellowish phyllosilicates prior to their deposition.

The age relationship of these deposits and the rest of the roadcut is questionable. Because very similar but more coherent rocks were found in other parts of the roadcut, these deposits are rather Tertiary in age than young, poorly consolidated Quaternary deposits. However, the deposits have a reduced coherence, probably due to weathering and decementation.

The deposits resulted from a mass flow that was competent enough to carry large clasts. The incomplete disintegration and incomplete mixing of large gray (volcanic rock clast dominated) blocks with reddish-brown (sedimentary rock-dominated) components indicate lack of turbulence. Disintegrated blocks of older deposits ("megaclasts", of Palmer et al., 1991; "debris-avalanche blocks" of Glicken, 1991) are typical of debris-avalanche deposits. Similar to avalanche deposits, the matrix here is mud-rich sand. The presence or lack of hummocky topography cannot be determined. The clasts do not seem to be "crackle" brecciated which would be an indication of an avalanche origin (Glicken, 1991; Lowe and Smith, 1991; Palmer et al., 1991). In spite
of some analogies to debris-avalanche deposits, most likely the deposition took place from a debris-flow.

Figure 25. Thick-bedded conglomerates, volcanic sandstones and poorly coherent conglomerates at Jordanelle (roadcut 2, Appendix 1). Poorly coherent breccias are massive beds with very large blocks of predepositionally altered volcaniclastic deposits (right). A 1 m thick wedge of clast-supported conglomerates is present to the right of the 1.5 m long Jacobs staff.
10.6.4 Thick-bedded conglomerates and volcanic sandstones

On the north side of the massive conglomerate deposits, thick-bedded polymictic conglomerates/breccias and volcanic sandstones can be found (Figures 25 and 26). In general, the conglomerates are clast-supported and inversely graded. Coarser, cobbly to bouldery beds are discontinuous and roughly lenticular. The thicknesses of inversely graded basal layers may reach 25-30 percent of the total thickness of the bed, and varies laterally over distances of several meters from vaguely developed to well-developed. Some of the coarse discontinuous lag-like beds lack inverse grading. In finer beds, the inversely graded portions may reach 50-70 percent of the total thickness. However, such beds have scattered over-sized cobbles and boulders that rest on the base. Some of the beds have slightly convex upwards surfaces and lenticular geometries. The beds grade laterally to volcanic sandstones. Volcanic sandstones contain dispersed pebbles and scattered large boulders. In places, the boundaries between sandstone beds are indicated by discontinuous strings of pebbles.

The polymictic clast compositions, clast-supported textures and inverse grading indicate that the coarse conglomerates/breccias were deposited by non-cohesive debris flows. The finer conglomerate beds with inverse grading involving most of the thickness are comparable to transitional flows of Scott (1988). However, here the deposits are coarser than those described by Scott (1988). In spite of slightly different grain-size distribution and possibly flow conditions, the mechanism of progressive deposition proposed by Scott is still probably applicable. As the mass flow was dilute, the faster moving fluidal head and body deposited the massive and poorly inversely graded parts which were then overlain by coarser and progressively thinner parts of the flows. Massive volcanic sandstone beds may represent more dilute flows in which coarse clasts had already been deposited and the entire flows had reached a hyperconcentrated state. Transitional and hyperconcentrated flows described by Scott lacked boulders. However, the flows described by Smith (1986) included boulders with long axis orientation perpendicular to flow direction. Such boulders were transported by rolling at the base and thus had a different transportation mode than the rest of the flow. Similarly, the over-sized boulders apparently resting on flow bases were probably transported by traction. Since the boulders are poorly to moderately rounded, they were not transported very far.
Figure 26. Thick-bedded conglomerates and volcanic sandstones showing a channel fill (center) at Jordanelle (roadcut 2, Appendix 1). The camera view is adjoining to the left of Figure 25. Note the grain size difference of the beds on opposite sides of the channel. Jacob's staff is 1.5 m long.
10.6.5 Channel-fill deposits

The deposits described above are in turn cut by another deep channel (Figures 26 and 27). The channel is filled by several depositional units.

At the base of the channel are stratified conglomerates containing subrounded or rounded polymictic clasts and breccia blocks (maximum Ø 50 cm). The breccia blocks are comprised of pumiceous lapilli tuff. These clasts indicate that "the light gray or white altered clasts" described in this work indeed include pumice as well as more massive volcanic rock fragments. Distinction of different clast types is difficult if alteration or weathering has been intense.

Two stratified conglomerate beds can be recognized. The upper is slightly finer and has a pinkish color. Stratified conglomerates are separated by discontinuous pinkish fine volcanic mudstone laminae that can be traced all the way up the steep slopes of the channel.

The stratified conglomerates are overlain by another discontinuous fine volcanic mudstone layer and a 10 cm thick massive volcanic pebbly sandstone containing altered clasts. Both layers are mantling channel floor.

The next bed is a 70 cm thick pebbly sandstone with scattered cobbles. The rest of the channel fill deposits are thick, inversely graded polymictic breccias separated by inversely to normally graded thin tuffaceous beds and fine tuffaceous laminae. These beds do not mantle the channel walls, but are instead horizontal.
Figure 27. Detail of the channel fill shown in previous figure. Scale division is 10 cm.
10.6.6 Interpretation of channel-fill deposits

Big blocks of pyroclastic rocks and well-rounded stratified conglomerates near the base of the channel suggest that the channel was cut by streams and that the first beds deposited in it were fluvial conglomerates and large blocks probably excavated from steepened walls. Subsequently it was filled with thick-bedded debris flows and hyperconcentrated flow deposits. Thin discontinuous volcanic mudstone laminae are too well-sorted and fine to be sole layers (Scott, 1988) of overlying deposits. Fine-grained deposits are formed in low-energy conditions and the depositional process could have been a pyroclastic fall or surge as well as a low velocity fluvial current.

10.6.7 Pumiceous stratified volcanic sandstones and massive breccias

The remainder of the roadcut comprises stratified pebbly volcanic sandstone and thick massive breccias. Volcanic sandstones (Figure 28) contain abundant pumice fragments. They are deposited on a partially exposed conglomerate that probably correlates with the thick bedded debris flow and hyperconcentrated flow deposits described previously. The first 0.7 - 1 m of the sequence (Figures 28 and 29) consists of distinctively stratified pumice-rich beds in which beds are thin to medium-thick.

The lowest sandstone bed is a thin massive fine volcanic sandstone layer which is followed by thin, crudely inversely graded, pebbly sandstone. The next bed is ca. 35 cm thick and shows peculiar lenticular grading in its upper part. The lowest 20 cm of the bed is a massive medium-coarse sand. The massive part grades into a lenticularly bedded part in which lenses up to 2 cm thick and several decimeters long contain abundant pumice clasts as well as a few andesite and crystal fragments (Figure 29). In the lenses, clasts are in the coarse sand or granule range. The bed is similar to some previously described deposits of outcrop 1B.

The next 30-45 cm consists of thin to medium-thick pumice-rich beds, thin medium-coarse sandstone beds, and thin pinkish-gray fine laminae. Pumice-rich beds are very thin lenses that continue laterally for 10-20 meters. The largest pumice clasts are in the small pebble range. Andesite clasts are granules. Pumice-rich beds seem to be associated with fine volcanic mudstone/fine sandstone layers at their bases. Laminae are discontinuous but commonly can be traced over 10 m. Their maximum thickness reaches 2 cm but typically the layer is only a few millimeters thick. Thicker parts
contain dispersed pumiceous granules and pebbles. A thin section from a 5 mm thick layer reveals well-sorted sandy siltstone with abundant broken crystal clasts (25-30 percent) and pumice (Figure 30). The clast composition of the bed above the sandy siltstone is listed in Table 7 (sample JOR RC2 2). Here, pumice contains quartz (and large biotite) phenocrysts, indicating a dacitic composition. Sparsely vesicular volcanic rock fragments include dacite clasts but plagioclase-amphibole porphyritic clasts are also present. The latter clasts include both plagioclase- and amphibole-dominating andesitic clasts. The texture of the groundmass is variable. In a few porphyritic clasts the originally glassy groundmass has devitrified to a chert-like aggregate. In a few clasts the groundmass is microcrystalline. The amount of various andesitic clasts is small and reworked origin cannot be established with certainty.

The medium-coarse sandstone beds intercalated with pumiceous sandstone and mudstone pairs show lenticular stratification similar to but slightly less distinctive than in the first unit.

Overlying deposits are characterized by thicker but less distinctive beds. The largest clasts are typically pebbles, but a few beds contain small cobbles. Beds generally contain altered light gray pebbles that are concentrated towards the bases (Figure 28). Commonly these clasts form clast-supported lenses, some of which are found above an inversely graded basal layer. Such clasts appear rather compact and probably they are not pumice. Many of the beds show inverse to normal or inverse grading. The inversely graded parts generally encompass most of the bed thicknesses and contain both altered clasts (low density pumice) and andesite clasts (Figure 31). Grading patterns typically persist laterally for 10-20 m. Bedding is not distinctive and beds appear to be subunits of thicker beds. Except for the light gray clast-rich lenses, the larger clasts are supported by a sandy matrix. The beds have altered light gray clasts that have a less coherent appearance. Such clasts also form lenses and stringers which, are commonly found at the tops of the bed. Thin section study verifies that such clasts are pumice. Pumice is also found abundant as sand- and granule-sized clasts in the matrix.
Figure 28. Pumiceous stratified volcanic sandstones in roadcut 2 (Appendix 1). The deposits rest on a clast-supported conglomerate. The lower 0.7 m consists of several thin beds. In the remaining part, the beds are thicker and less distinctively stratified. The scale is 1.5 m long.
Figure 29. Lenticularly bedded volcanic sandstones and thin pumiceous pebbly sandstone beds in roadcut 2 at Jordanelle (Appendix 1). The pumiceous beds at the top grade laterally into the pumiceous beds shown at the bottom of the previous figure. The pumiceous beds overlie 0.5 to 1 cm thick pinkish-gray laminae which also contain small pumice granules. Scale bar divisions are in 10 cm and 1 cm.
Figure 30. Photomicrograph of a volcanic sandy siltstone laminae separating pumiceous pebbly sandstones in roadcut 2 at Jordanelle (Appendix 1). The maximum thickness of the sandy siltstone layer is 8 mm.
Figure 31. Medium-thick to thick beds with diffuse to distinct contacts, scattered over-sized pebbles of sparsely vesicular andesite, and inverse grading in roadcut 2 at Jordanelle (Appendix 1). Note the inversely graded bed above the hammer.
10.6.8 Interpretation of pumiceous stratified volcanic sandstones and massive breccias

Because the clast composition is not distinctively polymictic, the depositional mode is debatable. Discontinuous beds do not favor pyroclastic fall as a depositional mechanism (Cas and Wright, 1987). Lenticularly bedded units found in the lower part of the sandstone sequence (Figure 29) are formed by tractional (i.e. clast-by-clast) deposition. The lenses resemble pumiceous discontinuous layers and lenses described by Smith (1986, figure 4). Here, however, the deposits are better sorted and finer grained. They also have relatively thick massive basal parts. Some graded stratified hyperconcentrated flows have massive basal parts but are normally graded. However, pumice lenses and concentration zones have been reported from "ash hurricane deposits" (Smith and Roobol, 1990) that probably represent intermediate conditions between pyroclastic surges and flows (Carey, 1991).

Thin pumice-rich beds do not show the well-developed grading pattern typical of pyroclastic flows. Fine volcanic sandstone and siltstone laminae could be ash cloud surge and fall deposits, but appear to be "ground" layers. They are too fine-grained for ground surges. Such beds could have been deposited by low-energy streams that occurred between the deposition of various coarse volcanioclastics.

Thick bedded pebbly sandstones are very similar to the deposits of transitional flows between debris flows and flows that had reached hyperconcentrated flow conditions (Scott, 1988). Transitional flows have inversely graded parts that involve most of the bed thickness, similar to the bed in Figure 31. The inverse grading involves both pumiceous and sparsely vesicular clasts. Scott has also described stringers of low density clasts towards the tops of the beds.

10.6.9 Massive channeled breccias

The volcanic breccias and sandstones are cut by thick massive breccias that are as thick as 15 m. There are three different depositional units filling deep channels.

A clast-supported massive breccia unit approximately 5 m thick (Figure 32) contains abundant large angular blocks and rounded boulders (Ø exceeding 1 m). The clasts are mostly sparsely vesicular dark grayish-andesites. However, the clast composition is polymictic. Many of the clasts have radial cooling-joints that indicate in situ cooling (Smith and Roobol, 1990; Smith and Lowe, 1991). Crackle-brecciated clasts were not
observed; instead, the breakage of the clasts can be almost exclusively attributed to tectonic jointing and faulting or blasting and excavation during road construction. The matrix is granular.

Polymictic breccias with radially jointed, "hot" clasts are also found at outcrop 8 (Appendix 1). Since the bedding is horizontal in outcrop 8 and roadcut 2, their deposits probably are contiguous.

The first "hot" breccia is overlain by another breccia containing in situ cooling brecciated clasts. The unit is exposed at the top of the roadcut and cannot be reached for detailed observations.

The third and thickest breccia is at least 15 m thick. It is cut by several faults which show normal dislocations of a few meters. The breccia unit contains blocks of older volcanic deposits as long as 5 m, including clast-supported breccias, altered and "slumped" phyllosilicate-rich blocks, and lithic arenite and conglomerate blocks (Figure 33). The blocks are coherent and lack disintegration shattering. The breccia also contains reddish patches and stringers that have irregular contacts. The reddish color probably results from alteration by fluids after deposition. However, the reddish patches and stringers contain more red sandstone fragments than the grayish parts of the the breccia (Figure 34). The alignment of the reddish stringers suggests partially preserved stratigraphy. Reddish sedimentary rock clast-rich material has been caught up in the flow and "swingled" a bit by the action within the flow. After deposition, the red sandstone fragments provided the source of the hematite stain causing the reddish color.
Figure 32. Stratified volcanic sandstones and overlying polymictic breccias in roadcut 2 at Jordanelle (Appendix 1). Two breccia units can be recognized. Large clasts have radial joint systems that indicate in situ cooling.
Figure 33. Large breccia block consisting of predepositionally altered and slumped volcaniclastic material and stratified volcanic sandstone in roadcut 2 (Appendix 1). Scale bar is 1.5 m long.
Figure 34. Sedimentary-fragment-rich reddish patches and stringers in gray breccia that is rich in volcanic rock fragments in roadcut 2 (Jordanelle).
10.6.10 Interpretation of massive channeled breccias

The polymictic composition and rounding suggest a reworked origin. The flows containing radially jointed clasts originated from transformations of pyroclastic flows to debris flows or perhaps more likely, by erosion of still hot pyroclastic material by snow melt, rain or streams (Smith and Lowe, 1991). The flow with large blocks of volcaniclastic rocks represents mass movement that redeposited volcaniclastic deposits around a volcano. The flow was not capable of completely mixing the debris. It probably resulted from a local slope failure of the channel wall (Rodolfo and Arguden, 1991; Smith and Lowe, 1991).

10.7 Roadcut 3

The lateral distance between roadcuts 2 and 3 is ca. 200 m. The covered interval between the massive debris flow deposits described in the preceding paragraph and the sequence of stratified volcanic sandstones (Figure 35) in roadcut 3 is ca. 10 m. Deposits are partly altered along bedding planes to yellowish phyllosilicates and replaced by black chert. In a 30 thick layer, the deposits have been altered to light gray clay except for small light brownish hydrobiotite flakes and some remnants of feldspar phenocrysts. The altered layer probably represents one medium-thick bed. Taking into account the signs of fault-related alteration in the roadcut, the alteration of this particular bed probably occurred after burial and therefore, the bed is not a paleosol.

Similar to the sample from roadcut 2, the sample from roadcut 3 consists of dacitic pumice and sparsely vesicular andesite. Most of the strongly altered light gray clasts are pumice. Beds are thick, commonly graded and lack fluid escape structures. Inverse grading may involve most of the bed thickness. Pumice clasts are poorly concentrated towards the tops. Lack of fluidization structures suggests a reworked origin. Grading patterns and grain-size distribution are similar to transitional and hyperconcentrated flow deposits of Scott (1988).
Figure 35. A strongly altered tuff layer in roadcut 3 (Appendix 1) is overlain by a ca. 1 m thick bed in which the uppermost 20 cm is pumice-rich. The succeeding bed is a lapilli tuff showing inverse to normal grading of dense clasts and having a laterally discontinuous pumice lapilli-rich top division, 5 cm thick, near the upper end of the 1.5 m long scale bar.
10.8 Roadcuts 4 and 5

10.8.1 Stratified volcanic breccias and sandstones

The next stratigraphically higher deposits (roadcuts 4 and 5) are stratified volcanic sandstones, breccias and conglomerates forming a 12 m thick sequence. Roadcut 4 is ca. 100 m to the north of roadcut 3. The covered stratigraphic interval between the roadcuts is less than 10 m. The lateral distance between roadcuts 4 and 5 is ca. 70 m.

Pebbly and cobbly beds are medium-thick to thick. Laminated sand and gravel beds are thin and somewhat less continuous than the coarse ones. In general, the clasts are subangular to subrounded. The clast composition is polymictic; grayish plagioclase-amphibole-porphyritic andesites dominate. In addition to various types of volcanic rock fragments, accessory pebbles and small cobbles include quartzite and gneiss. Coarse beds are clast-supported and contain scattered boulders. In at least the lower 5 m (Figure 36), the beds are massive or show grading that is either crude inverse, inverse to normal, or normal. The beds fill gentle channels and can be traced laterally for 5-10 m. Bedding contacts vary from relatively gradual to sharp.
Figure 36. Stratified polymictic pebbly sandstones and conglomerates in roadcut 4 (Appendix 1). Scale bar is 1.5 m long.
The crudely stratified deposits are followed by a 7 m thick sequence in which stratification is more distinct (Figures 37 and 38). The lower portion is coarse and thick bedded but the upper 4 m are comprised of pebbly and sandy deposits. The coarse beds are inversely or inversely to normally graded. The inverse grading involves 30 percent or less of the bed thickness. Typically, the average grain size is in the cobble range. The beds contain few boulders that protrude into overlying deposits. Bases of the beds are clearly erosional or sharp. However, some repeated inversely to normally graded beds give an impression of pulsating deposition that resulted in gradual contacts. The matrix is sandy and contains small strongly altered light gray clasts that are not found as larger clasts.

Thin sandstone interbeds are massive or crudely horizontally laminated. Thicker beds in the upper part of the sequence are highly weathered but appear to be horizontally stratified. The stratification is laterally continuous but is locally gently channeled. The beds contain cobbles and clast-supported cobbly lenses.

10.8.2 Interpretation of roadcuts 4 and 5

Observed crude stratification and grading patterns suggest a high clast concentration prior to deposition and "flashy" flow conditions approaching those of hyperconcentrated flows (Nemec and Steel, 1984; Smith and Lowe, 1991). Inversely graded and inversely to normally graded coarse deposits are deposited by fluid-rich debris flows in which basal shearing was effective. These flows were competent enough to carry boulders well above their base or even partially protruding from the flow. Normally graded parts were formed in less competent debris flows in which large clasts had started to settle differentially (Scott, 1986, 1988). Thin crudely stratified and thicker stratified sandstone beds were formed by hyperconcentrated flows segregated from debris flows or by more dilute lahar-related stream flows.

The intensity of alteration increases laterally to the north. In the middle of an approximately 100 m wide zone is a spheroidally weathered or altered plagioclase porphyritic intrusion. This 10 m wide body may be related to the Jordanelle Intrusion (Woodfill, 1972; Bryan, 1990) exposed in the Provo River Valley 400-500 m west of the roadcut. Otherwise in this zone, all primary structures have been destroyed.
Figure 37. Polymictic breccias and volcanic sandstones in roadcut 5. Scale bar is 1.5 m long.
Figure 38. Detail of the same deposits shown in previous figure. Scale bar is 1.5 m long.
10.9 Roadcut 6 (Jordanelle dam over-look area)

10.9.1 General description

A sequence of breccias, conglomerates and interbedded volcanic sandstones at least 8 m thick is well exposed on the roadcut at the Jordanelle dam over-look area (Appendix 1). The exposure is about 100 m wide but at its southern end the deposits are intensively altered. The deposits comprise a strongly channeled sequence and several lithologies can be recognized.

10.9.2 Description and interpretation of roadcut 6

Clast-supported coarse deposits vary greatly in thickness, texture and grain size. Cobbly and bouldery conglomerates occur as a few-clast-thick lags in which coarse clasts were left resting on an erosional surface and were finally mantled under finer deposits. Some of the clast-supported deposits contain rounded cobbles and pebbles in a sandy matrix. Such deposits could represent fluidal mass flow or stream flow deposits.

Several meter thick massive, very poorly rounded breccias fill deep channels. Studied deposits contain large blocks, the diameters of which exceeds 2.5 m. Some of these units show crude bedding with gradationally alternating coarse pebble and cobble beds. These beds are likely fluidal debris flow deposits, although crude bedding may result from a surging flow (Nemec and Steel, 1984).

Matrix-supported conglomerates and breccias contain large blocks and boulders. Thick beds have sharp or erosional bases and a muddy sand matrix. Clast content is high and the texture is nearly clast-supported. Matrix-supported breccias and conglomerates are deposited by debris flows in which cohesion was an important clast-support mechanism.

Volcanic sandstones include thin horizontally or small-scale cross-laminated beds overlying clast-supported conglomerates. Horizontally laminated sandstones may result from hyperconcentrated flows associated with debris flows or from more dilute high-velocity stream flows. Cross-laminated beds are deposited from stream flows (Allen, 1982; Collinson, 1986; Nemec and Steel, 1984; Smith and Lowe, 1991). Some very thin to thick sandstone beds are normally graded and also contain over-sized pebbles at their bases. Grading is a well-developed distribution grading. The clast composition of
a normally graded bed is listed in Table 7 (sample JOR RC6 1). An inversely graded bed 30 cm is present overlying a conglomerate lag. The entire thickness of the bed is graded. Normally graded sandstones required a high sediment concentration, such as in hyperconcentrated flows (Lowe, 1982; Nemec and Steel, 1984; Smith and Lowe, 1991). Inverse grading that is not restricted to the basal part of a flow can be formed by progressive deposition of a vertically or laterally graded flow. Alternatively, the boundary of the flow plug may migrate due to changes in slope or in the roughness of the base. Because inversely graded sandstone is overlying a channel lag conglomerate, the latter explanation is more probable.

10.10 Roadcuts 7-9

On the following two roadcuts (roadcuts 7 and 8, Appendix 1), deposits are altered along a conjugate joint system and this obscures observations. The deposits are massive breccias in which the diameters of the largest blocks exceed 1 m, but the average grain size is 30-40 cm. Their compositions vary from monomictic to polymictic. The former breccias characteristically include plagioclase-amphibole-porphyritic volcanic rock fragments with scarce or no biotite phenocrysts, or dark brownish plagioclase porphyritic clasts with an aphanitic groundmass. Polymictic breccias contain reddish-gray plagioclase porphyritic clasts with aphanitic groundmass and additional pyroxene-amphibole porphyritic mafic lava fragments and plagioclase-biotite porphyritic clasts. Textures vary from matrix-supported to clast-supported.

The last studied roadcut along US 189 (roadcut 9) provides an impressive, several tens of meters thick cross-section of massive and strongly altered coarse breccias. Altered breccias are cut by unaltered to poorly altered reddish-gray plagioclase-amphibole-porphyritic intrusions and dark bluish-gray amphibole porphyritic dikes. Breccias appear to be massive. However, perhaps because of alteration, bedding contacts were not observed. In places the breccia is clast-supported but in other places appears to be matrix-supported.

10.11 Outcrops 8-21

In outcrops 8-21, the rocks are andesitic breccias with a few basaltic lava flows and andesitic tuffaceous interbeds. One lava flow has a grayish-red top (outcrop 21,
Appendix 1). In addition to oxidation, proximity of a vent is indicated by partly agglutinated bombs. (The petrography and probable origin of the horizontally stratified scoriaceous deposits associated with these lava flows were briefly discussed in Chapter 9.) Volcaniclastics here are cut by dikes and shallow intrusions. Coarse volcanics fill deep channels. Complicating the stratigraphy are apparently normal faults with dislocations of tens of meters. Taking this into account, the following stratigraphic sequence is tentative and is inferred from the elevations of each outcrop.

At outcrops 8 and 9, volcanic breccias are polymictic, containing plagioclase-amphibole porphyritic andesite clasts. Breccia clasts differ in phenocryst content and the degree of crystallinity of the groundmass. Dark gray andesite clasts are pieces of cooling-jointed lava. Amphibole phenocrysts are clearly hornblende, and deviate thus from andesitic clasts that usually contain oxyhornblende. Polymictic compositions, slight rounding of some of the clasts, the massive appearance, and the muddy sand matrix suggests that these breccias are debris flow deposits.

At outcrops 10 and 11, the breccias are more monomictic. Breccia clasts are plagioclase porphyritic andesites with accessory amphibole and clinopyroxene phenocrysts. The relatively glassy groundmass contains abundant small feldspar (plagioclase and sanidine) laths and small accessory orthopyroxene, biotite and opaque crystals.

The breccia matrix is coarse sand with a few percent clay (mostly probably diagenetic) and hematite cement. The shapes of the clasts suggest fragmentation by mechanical breakage rather than explosive shattering or quenching. The breccia clasts have mineralogies and textures similar to some of the minor shallow intrusions found in the area. The breccias may be redeposited or block and ash flow deposits resulting from the collapse of a dome.

The monomictic breccias are covered by a thick sequence of polymictic clast- to matrix-supported breccias (outcrops 12 and 13). At a 6 m thick outcrop (# 13), no bedding was observed. The breccia matrix is muddy sand with abundant granules and small pebbles. Large blocks are 0.5-1 m in diameter, although breccia clasts are typically a few decimeters in diameter. The largest clasts are commonly in contact with each other while smaller pebbles and cobbles are surrounded by granular matrix.

Above the polymictic breccia, a horizon of plagioclase-pyroxene-amphibole porphyritic basaltic-andesite rubble is present (outcrop 15). The clasts are highly
vesicular and the horizon represents aa-lava. It is in turn overlain by a breccia that contains clasts of underlying deposits, including fragments of underlying aa-lava. Therefore, the deposits clearly have a reworked origin.

The rest of the studied outcrops include dikes (outcrops 17, 18) and possibly primary pyroclastic breccias containing abundant pumiceous, strongly altered clasts.

11. Francis and Provo River Valley

Some further mapping was done at an old quarry near Francis (Francis Quadrangle, the northwest corner of section 32, T.2 S. and R.6 E.) and at roadcuts in the Northern Provo River Valley (Francis Quadrangle, the northern halves of sections 34 and 36, T.2 S. and R.5 E.).

At the outcrop near Francis (Appendix 4) an approximately 20 m thick conglomerate sequence is overlain by approximately a 30 m thick sequence of tuffaceous volcanic breccias. The conglomerate sequence is probably fluvial but access to these outcrops was prohibited by steep slopes, vegetation and bellicose steers.

Overlying tuffaceous volcanic breccias contain approximately 30 percent lithic fragments (Ø 3-7 cm). Scattered blocks are up to 35 cm in diameter. All lithic fragments contain plagioclase and amphibole phenocrysts (Ø up to 3 mm). The amount of clinopyroxene phenocrysts varies from trace amounts to 8 percent. Also one lithic sample contained pigeonitic clinopyroxene (as suggested by the small 2V value, Tröger, 1952). Another sample has agglutinated fragments. Some samples are altered to a yellow phyllosilicate. The pumice clasts and the highly vesicular clasts comprise 20 percent of the rock. The grain-size distribution of pumiceous clasts seems to be bimodal. Larger pumice or highly vesicular clasts are 2-6 cm in diameter. Smaller pumice clasts (Ø 0.4-1.5 cm) are scattered through the tuffaceous matrix. The alteration of pumiceous clasts to clays is not as strong as at other studied outcrops. The matrix is gray muddy sand.

The breccias are capped by over 20 m of sparsely vesicular andesite. This andesite lacks vesicle-rich or brecciated horizons. Instead of being a lava flow (Woodfill, 1972) it probably is a sill or a shallow intrusion.

In spite of approximately horizontal bedding at Francis, similar deposits were not found at Provo River valley. These deposits are characterized by shallow intrusions and
lobate breccias tens of meters thick. Some of the breccias are very monomictic with an aphanitic compact matrix. Such breccias may be block lava flows gradational to the shallow intrusions. Some of the polymictic breccias contain clasts with large plagioclase phenocrysts. In some of the clasts, plagioclase phenocrysts are over 1.5 cm long whereas the groundmass is aphanitic. Such large phenocrysts were not found at other studied outcrops. The roadcuts of the Provo River Valley were studied very briefly and conclusions as to their depositional mechanisms are not offered. Nevertheless it is clear that the breccias of the Provo River Valley were derived from a source that was not contributing material to the other studied deposits.
12. Deposits at Indian Hollow

The studied outcrops at Indian Hollow are located in the Kamas Quadrangle, in sections 11-15 of T.2 S. and R.5 E (Appendix 2).

Indian Hollow is a valley that is open to the southeast. At the northwest edge of the valley, a shallow intrusion (Indian Hollow plug, Woodfill, 1972) and a radial dike system are exposed. A 50 m thick sequence of stratified volcanic breccias with minor volcanic sandstone interbeds is exposed along the northern fork of a creek flowing along the eastern edge of Indian Hollow. Above this sequence, basaltic scoria breccias, lava flows and interflow sandstones are exposed. In the following pages, these deposits are referred to as a lava-dominated sequence. Its total thickness is approximately 100 m. Some outcrops where studied at the southern fork of the creek near the northwest edge.

12.1 Volcanic breccia-dominated sequence

The characteristics of the coarse deposits were not always distinctive. The differences between matrix-supported and clast-supported texture are not easy to define if the matrix is granular. Also, differences between breccias and conglomerates are not definitive. Therefore, the following lithologies are based on a loose set of characteristics.

13.1.1 Coarse breccias

Coarse breccias can be up to 4 m thick, but typically beds are 1-2 m thick (Figure 39). Beds contain scattered large blocks which are as large as several meters in diameter. In general, breccia clasts are in the cobble or boulder range. The distinction between matrix- or clast-supported texture is almost arbitrary. The matrix is coarse and contains abundant granules and small pebbles. Most of the granule- and pebble-sized clasts are dense or moderately vesicular. However, the matrix contains sand- and gravel-sized pumice or highly vesicular clasts (ca 40 percent, Table 7, sample IH 5). The breccia clasts are typically subangular or subrounded and overwhelmingly sparsely vesicular or dense. In most beds many large blocks support each other, but smaller (cobble-sized) clasts appear to be surrounded by coarse granular matrix. Small block- and boulder-sized clasts are commonly protruding into overlying deposits. Larger clasts
are generally within the bed or at the base. However, matrix-supported beds may grade laterally into clast-supported beds. Some breccia beds show inverse grading, with the matrix-supported pebbly basal division several decimeters thick. Bases above finer deposits are sharp or even erosional but in cycles of thick and coarse breccias, the contacts can be defined only with difficulty. However, some of the otherwise poorly recognizable depositional units are segregated by thin pinkish volcanic sandstone/siltstone laminae.

In some beds the matrix is reddish-gray. As in breccia units with gray matrix, clast composition appears to be polymictic. Breccia clasts are sparsely vesicular or dense andesites but differ slightly in phenocryst content and size as well as by the color and crystallinity of their groundmass. Some of the blocks with a brownish weathering surface apparently have a microcrystalline rather than an aphanitic groundmass. Clasts rarely show radial, \textit{in situ} cooling-jointing. Instead, blocks commonly seem to have facets with $120^\circ$ angles. These are probably pieces of previously cooled jointed rocks. Some beds are dominated by large stratified volcanic sandstone blocks (Figure 40).

Another example of a breccia unit with a distinctively reddish matrix is present at outcrop 5 (Figure 40). The unit is clearly matrix-supported. The maximum-clast size is 80 cm. The clasts are sparsely vesicular plagioclase-amphibole porphyritic andesites with glassy matrices. Some of the large clasts display \textit{in situ} cooling-joints. One bed contains small reddish andesite clasts in which the amount and size of plagioclase phenocrysts is similar to larger clasts. The amount of reddish clasts is higher in the reddish layer than in the overlying grayish one. There are no grains that are only partially reddish or that have only reddish stain on the surface. Although deposits can derive a red color from diagenetic oxidation, in this case the reddish color is undoubtedly due to primary oxidation. The reddish color and \textit{in situ} cooling-jointed clasts indicate that the matrix of this particular bed was hot at least during transportation and possibly, similarly to larger blocks, cooled after deposition. With its monomictic breccia clast composition, oxidized matrix and distinctive matrix-supported texture, the unit deviates from other observed breccias.
Figure 39. Thick, coarse breccias at Indian Hollow with finer (cobbly to pebbly) breccia lenses to the right of the hammer (outcrop 8, Appendix 2).
Figure 40. Matrix-supported coarse monomictic breccia with reddish-gray matrix and scattered cooling-jointed blocks (outcrop 5 at Indian Hollow, Appendix 2). The overlying breccia approaches a clast-supported texture and contains large blocks of stratified volcanic sandstone.
13.1.2 Finer breccias

Finer breccia beds generally have variable clast compositions, but the breccias of outcrop 3 have a relatively uniform clast composition (Figure 41). Altered light gray clasts are 0.5-1 cm in diameter. Many of them are probably pumice. Most andesite clasts are in the pebble or cobble range, but some are a few decimeters long. A few clasts have radial cooling-joints. Andesite clasts are moderately or highly vesicular and plagioclase and amphibole porphyritic (approximately 20 percent and 10 percent respectively). Clinopyroxene and biotite phenocrysts are present in accessory amounts. The majority of the clasts are dark bluish-gray and have a glassy groundmass. Slightly reddish clasts are similar to dark bluish ones except for oxyhornblende phenocrysts and small opaque crystals in glassy groundmass; the latter crystals were probably derived from the oxidation of amphibole crystallites to opaque. Grayish clasts have an altered (palagonitized) groundmass. Some of the breccia clasts are flow-banded and consist of both grayish and dark bluish parts. The breccia "matrix" contains abundant pumice as granule- and sand-sized grains (30 percent, Table 7, sample IH 3). Most of the clasts are sand-sized rock fragments and broken crystals. Andesite clasts constitute 41 percent of the sample. Their vesicularity varies from dense to sparsely vesicular. Due to alteration, clast contacts are not easily distinguishable. Despite this, shards representing broken bubble walls are found as small clasts. Less than 10 percent of the sample is matrix; taking into account the alteration of pumice clasts, most of it is probably diagenetic (Dickinson, 1970). The over-all impression is that the great majority of clasts are derived from the same source and that their differences in color and mineralogy are due to different oxidation and alteration stages.

Some beds are dominated by dark bluish-gray andesite clasts, but also contain other clasts suggesting a reworked origin. An example of such a bed is shown in Figure 42. Most of the clasts are plagioclase-amphibole porphyritic andesites with dark bluish-gray glassy groundmasses. However, clasts also include plagioclase-amphibole-biotite porphyritic andesites and clasts that are pieces of lavas that were jointed in a columnar manner. In many beds, the clasts are polymictic.

The textures of finer beds can vary from distinctively clast-supported to matrix-supported. In many beds, the texture is better described as matrix-supported with a coarse granular matrix. Internal structures include inverse grading and crude stratification.
Figure 41. Relatively monomictic pebble and cobble breccias (outcrop 3, Indian Hollow, Appendix 2). Textures vary from distinctively matrix-supported (the bed below the hammer) to clast-supported (the partly visible bed with the erosional base at the top of the photo). Both bed types can have inversely graded bases.
Figure 42. Detail of a clast-supported bed at Indian Hollow (outcrop 5, Appendix 2). Clasts are dominated by plagioclase-amphibole porphyritic dark bluish-gray andesite clasts, with a glassy groundmass. The clasts include minor to accessory amounts of other lithologies such as reddish plagioclase-amphibole-biotite-porphyritic andesite clasts. Some clasts, such as the clast at the center, are pieces of columnar jointed lava. The overlying bed is inversely graded with a pebble-rich basal layer.
Clast-supported breccias are thick to very thick (<1.5 m). The bases are typically sharp and channeled. Some beds have scattered oversized blocks or boulders. The average grain size is pebble or small cobble. The beds can be crudely to distinctively inversely graded. The clast composition is variable. An example of a nongraded polymictic breccia is shown in Figure 39. An example of an inversely graded bed with a much more uniform clast composition is shown in Figure 41.

Medium-thick to thick, matrix-supported breccias (Figure 43) also contain oversized clasts that may protrude into overlying beds. As the clast content increases, matrix-supported beds gradationally change to clast-supported. Large clasts may form clusters or short stringers. The beds have thin inversely graded basal layers. The matrix is pumiceous sand with abundant fine gravel-sized clasts.

13.1.3 Crudely stratified breccias

Breccias can show a crude stratification (Figure 44). The matrix-supported breccias contain a few decimeter thick, laterally discontinuous, nearly clast-supported layers. The contacts of these layers are clearly gradational. Larger clasts are sparsely vesicular plagioclase- amphibole porphyritic andesites. Matrix-supported parts contain clusters of larger clasts. The matrix is sandy and contains small lithic and pumiceous clasts (Ø 0.5-1 cm).

13.1.4 Conglomerates

Especially in the upper half of the breccia-dominated sequence (outcrops 8-10), the coarse beds contain better rounded clasts. Many depositional units contain enough rounded clasts to be called conglomerates. The coarse (bouldery) conglomerates are both matrix- and clast-supported, and as thick as 2.5 m. Finer conglomerates are typically clast-supported. Distinctively clast-supported pebbly or cobbly conglomerates are comparable in composition to the clast-supported breccias described above. The coarser, bouldery conglomerates commonly have inversely graded bases similar to the coarse breccias.
Figure 43. Clast-supported (a), clast-supported and inversely graded (b), and matrix-supported breccias (c) with a thin massive sandstone bed (d) at outcrop 2, Indian Hollow (Appendix 2). The breccia clasts vary in mineralogy and texture. Some breccia clasts show in situ cooling-jointing (ISCJ); this jointing formed in hot breccia clasts as they cooled in the flow. Scale bar is 1.5 m long.
Figure 44. A crudely stratified matrix-supported breccia (behind the 1.5 long scale bar) overlain by a pumiceous breccia 30 cm thick and then ca. 30 cm thick clast-supported breccia with erosional base (outcrop 1, Indian Hollow, Appendix 2). Distinctive "layers" near the top of the pumiceous breccia are shear planes. Shearing took place subparallel to bedding and caused apparent dislocation of a few centimeters.
13.1.5 Pumiceous breccias

The breccias and conglomerates containing abundant sparsely vesicular andesite clasts are more resistant to weathering than are basal pebbly sandstone layers of inversely graded coarse breccias and conglomerates, volcanic sandstone beds, or breccias in which breccia clasts are dominantly altered light gray clasts (Figure 45). Some of these clasts are pumice, as verified by thin sections. The larger clasts are probably fragments of pumiceous tuffs.

Weathered, altered grayish-white clast-rich breccia beds are medium-thick or thick. The bases may be slightly erosional. Some of the beds may be inversely graded at their base. Sparsely vesicular clasts are considerably smaller (Ø in centimeters) than strongly altered pumiceous clasts (Ø in decimeters). An example of a nongraded bed is present at outcrop 5 (Figure 45). This altered light gray clast-rich bed is almost 2 m thick. The bed is matrix-supported approaching a clast-supported texture. The matrix is mud-poor and granular. Here most of the altered clasts are pumice. Even the largest clasts (Ø 80 cm) are highly vesicular. The lower boundary is sharp.

A thin pumiceous breccia bed is shown in Figure 46. This particular bed (outcrop 4, Appendix 2) is inversely graded and overlies basal laminae a few millimeters thick. In the major portion of the flow, sparsely vesicular clasts and crystal fragments are normally graded while larger pumiceous clast are found near the top.
Figure 45. Pumiceous breccias. The large altered light gray clasts show a weak layering and a "spotted" texture; those clasts are probably pieces of pumiceous tuffs. The upper breccia bed shows inverse grading that involves also scattered sparsely vesicular clasts (southeast edge of outcrop 5, Indian Hollow, Appendix 2).
Figure 46. Thin pumiceous breccia. The bed overlies a few millimeter thick fine laminae and has a 1-1.5 cm thick inversely graded basal part. Pumiceous clasts are concentrated at the top, whereas coarse matrix grades normally (outcrop 4, Indian Hollow, Appendix 2).
13.1.6 Volcanic sandstones

Two crudely stratified beds of volcanic sandstone were observed (outcrops 1 and 8, Appendix 2). The beds are thin to medium-thick, and contain a few dispersed sparsely vesicular andesite pebbles. Massive nongraded pebbly sandstone beds are medium-thick to thick. The overall grain size is coarse sand with dispersed granule and small pebble-sized red and dark gray, glassy andesite clasts. Some massive nongraded beds also contain altered light gray pebbles. They probably represent a finer variation of the above-mentioned pumiceous breccias. Such beds may have a poorly developed inversely graded basal layer. Massive volcanic sandstones with only scattered pebbles are thin bedded.

Massive beds with poorly developed thin inversely graded basal layers are probably gradational to more distinctively inversely graded pebbly sandstones (Figure 47). Such beds (3 observations) are thin to medium-thick. In one of the pebbly sandstone beds, dark gray or dark reddish-gray andesite pebbles are concentrated in the lower part of the flow but above a few centimeters thick basal layer. Thus the bed shows inverse to normal grading. Another bed (Figure 47) has a 40 cm wide normally graded pebbly scour-and-fill structure that grades laterally to volcanic sandstone. The sample of volcanic sandstone has a polymictic clast composition (Table 7, sample IH 7). In this scour-and-fill structure, the pebbles are subrounded to rounded. Another medium-thick massive sandstone bed at the same outcrop contains a few pebbles near its base. Apparently normally graded units are gradational to massive volcanic sandstones.

Fine volcanic sandstone or siltstone laminae also occur between coarse breccias and pebbly volcanic sandstones. Some of these laminae contain small pumice lapilli.
Figure 47. Volcanic sandstone and pebbly sandstone beds separated by volcanic siltstone laminae (a) in outcrop 7, Indian Hollow (Appendix 2). Sandstones include massive (b) and pebbly lithologies (c). Graded pebbly sandstones show inverse (d) and normal grading (e). Side of the plate is 7.6 cm (3") long.
12.2 Vertical changes in lithology

At outcrops 1 and 2 (Appendix 2), bedding is dipping toward the southeast. Since the dip is steeper than the river gradient, these deposits are stratigraphically lower than those exposed down-current in the gully (Outcrops 3 and 5-10, Appendix 2). Outcrops 8-10 may represent approximately the same stratigraphical level.

Breccias and volcanic sandstones are exposed at the hill top 6848 ft (Appendix 2) suggesting that the breccia sequence is more than 80 m thick. The deposits of outcrop 2 (Appendix 2) contain a few centimeter-wide subvertical mylonite zones as well as faults that are subparallel to bedding. The observed maximum dislocation of the latter type of fault is one meter. Field observations in the surroundings of hill 6848 ft suggest that some of rock units have been vertically displaced tens of meters. Therefore deposits at hill 6848 ft cannot be reliably correlated with other outcrops.

Coarse breccias with a reddish-gray matrix occur in the lower half of the breccia-dominated sequence. Similarly, finer breccias containing predominantly andesite clasts with dark bluish vitric groundmass were found only in the lower portions, although some upper polymictic breccias include such clasts. In the upper half of the sequence, deeply weathered layers are thinner and do not include pumiceous breccias. The uppermost breccias (outcrops 7-10, Appendix 2) do not contain very large blocks (Ø > 1 m) and some coarse deposits appear slightly better rounded.

12.3 Interpretation of the volcanic breccia-dominated sequence

In pyroclastic flows the temperature of the pyroclastic material commonly exceeds the Curie temperatures of the ferromagnetic minerals commonly present in trace amounts in volcanic rocks (such as titanomagnetite with the Curie temperature ca. 600°C). Therefore, the cooled deposits commonly show isotropic paleomagnetic properties. In reworking sedimentary mass flows the intergranular fluid is water and the temperatures are commonly lower (except for the possible hot breccia clasts). Therefore, reworking commonly leads to anisotropic properties. Because paleomagnetic measurements were not done during this study, it is difficult to define the origin of coarse matrix-supported breccias. Erosional bases indicate a flow origin. Coarse matrix-supported breccias lack convincing evidence of high temperatures during deposition. Slightly varying lithologies of clasts and the fact that many of them are
pieces of previously fractured rocks, suggest an origin as redeposited pyroclastic material. The most probable redepositing process for the breccias with a relatively granular matrix is deposition by competent debris flows in which cohesion did not play a very significant role as a clast-support mechanism. However the presence of some in situ fractured clasts suggests that still-hot debris was incorporated into the flows. This could have happened either by mixing of a pyroclastic flow with a stream or snowmelt water, or by bulking of stream or snowmelt water with still-hot pyroclastic material soon after the material came to rest (Pierson, 1985; Smith and Lowe, 1991).

A pyroclastic origin cannot be completely rejected. Pyroclastic flow deposits often contain accessory and accidental lithic clasts. Breccias resulting from violent ignimbrite eruptions may contain significant blocks from various previously existing rocks. A pyroclastic origin is even most probable for the matrix-supported breccia with a reddish-gray matrix and monomictic, partly in situ cooling-jointed blocks. Such breccia might be a proximal ignimbrite or perhaps more likely a block and ash flow deposited after an explosive dome collapse created some pumiceous ash. Laterally continuous inversely graded beds showing an abrupt increase in grain size resemble lag fall breccias (Walker, 1985; Cas and Wright, 1987, pages 237-242 and 482). However, single depositional units are much thinner, and their inversely graded basal layers and matrices contain abundant sand-and ash- sized clasts. Breccia layers are not overlain by thick ignimbrites or welded deposits. Therefore the coarse breccias found in Indian Hollow are probably not proximal co-ignimbrite breccias (i.e. lag fall, groundlayer or lithic concentration zone breccias) (Walker, 1985, Cas and Wright, 1987, p. 482). If not capped by welded deposits, such deposits have a poor preservation potential in the stratigraphic record (Cas and Wright, 1987). The lower the preservation potential, the higher the potential to be reworked. Such proximal deposits may have provided the source for most of the coarse breccias found in Indian Hollow.

The pebbly to cobbly breccias also lack gas escape structures and distinctive low and high density clast-enrichment zones. These deposits probably represent debris flows. Cohesion may have been significant in matrix-supported, massive and crudely stratified deposits. Crude stratification may indicate surging or pulsating flow, similar to some debris flows described in non-volcanic environments (Nemec and Steel, 1984). Some of the pumiceous clast-rich breccias are inversely graded, but sparsely vesicular clasts did not show normal grading. The thin bed shown in Figure 46 is a unique exception.
Some of the stratified and thin massive volcanic sandstone beds resemble sole layers of debris flows described by Scott (1988). Many of the thicker sandstone units are separated from breccias and other sandy deposits by fine volcanic sandstone or siltstone laminae. They represent separate depositional units. Volcanic sandstone beds probably represent debris flow deposits; especially, the normally graded beds with a polymictic clast composition are likely hyperconcentrated flow deposits.

Fine sandstone and siltstone laminae were likely deposited from ash clouds regularly occurring in this volcanic environment.

Therefore, as a whole, the breccia-dominated sequence was mostly deposited by debris flows, although some of the monomictic breccias in the lower half of this 50 m thick sequence have been deposited by pyroclastic flows. The uppermost 20-30 m are more clearly reworked deposits.

12.4 Interpretation of lava-dominated sequence

The contact between the breccia-dominated sequence and the overlying lava-flow dominated sequences is not exposed. On the western slope of hill 6961 (Appendix 2), the breccia blocks protrude through a thin soil and lichen at elevation of 6600'. Above this elevation, the slope is steeper. Loose boulders and blocks found in the soil are dark grayish andesite. At the 6640-6680' level, to the southwest of the hill top, a peculiar trace of Nugget Sandstone clasts is present. Some of these sandstone clasts appear to be pieces of tectonic breccia. Angular fragments are also present as xenoliths in lava-blocks.

Outcrops 15 and 17 above stratified volcanic breccias are dark lilac gray scoria breccia. Clasts are plagioclase-amphibole-pyroxene porphyritic andesite. Reddish clasts appear to be slightly more vesicular, which suggests that these clasts were primarily oxidized. A few less vesicular blocks are also present. The clasts are not welded, but many of them are slightly flattened and also have a poorly developed preferred orientation with subvertical long axes. The clasts are held tightly together by a dark grayish-lilac aphanitic and relatively compact matrix. The homogeneous composition and the texture leaves little doubt that these deposits are pyroclastic. The subvertical orientation of clasts and the presence of fine matrix (indicating moderate or poor sorting) suggest a pyroclastic flow origin.
Another andesitic breccia was found at outcrop 12. The andesite blocks are vesicular and plagioclase-pyroxene-amphibole-porphyritic. The matrix is fine-grained and tuffaceous. The origin of these deposits cannot be interpreted from such a small exposure, but the homogeneous composition and poor sorting suggest a pyroclastic flow.

The more than 7 m thick breccia sequence of outcrop 18 has a different appearance. The deposits are crudely stratified (0.5-1 m) breccias. Large volcanic-blocks are found at the tops of the beds. A few vesicular blocks seem to contain cooling-joints. Most of the finer breccia clasts are variably altered scoria with a composition (pyroxene-amphibole porphyritic basaltic-andesitic) similar to the larger clasts. Reddish-gray, dense to sparsely vesicular and light grayish vesicular plagioclase-amphibole porphyritic andesites are also present (<10 percent).

The intergranular matrix sample comprises scoria fragments (Ø 0.2-1 cm). Diagenesis has caused the the filling of vesicles by zeolites and later by fibrous clay. Fibrous clay is also found locally as a cement. The most prominent diagenetic product is clayey matrix. Glassy parts are strongly altered to a brownish zeolite-clay mesh, making the distinction of clast boundaries very difficult. The phenocrysts are mostly augite. Some brown oxyhornblende is present, as well as a few remnants of mafic phenocrysts that are completely altered to a rusty brown mesh (iddingsite?).

The slightly polymictic clast composition and the variable alteration of clasts suggests a reworked origin. However, the amount of minor clasts is small and the variable degree of alteration may be postdepositional and depend upon variations in vesicularity.

Clearly reworked deposits are present at outcrop 14 (Appendix 2). The attitude of the bedding is close to the outcrop surface. Bedding is dipping gently (26°) to the northwest. The stratigraphic thickness of these deposits is approximately 6 m. The lowest 1.5 m is clast-supported breccia in which most clasts are vesicular amphibole-pyroxene porphyritic andesite. Other deposits comprise 2 to 5 decimeter thick channeled clast-supported cobbly conglomerates with some small boulders (maximum Ø 30 cm) and 10 to 15 cm thick pebbly sandstone or fine conglomerate interbeds. The clasts are subrounded or rounded. The low-angle trough cross-bedding (formed by migrating channels), the polymictic composition, and the rounded clasts indicate traction sedimentation from high velocity streams.
The remainder of hill 6961 outcrops are lava-flows. At outcrop 19, a coarse andesitic aa-type lava flow is cut by a 20 m thick dike of plagioclase-hornblende porphyritic andesite. Lava-breccia clasts are dark brownish-gray andesite with 20 percent amphibole and pyroxene phenocrysts as long as 6 mm.

The breccia at the north end of the outcrop appears to be more blocky. Less vesicular blocks (maximum Ø 30 cm) grade to coarse granular matrix, apparently without gaps in grain size, giving an impression of a block-lava flow.

The lowest lavas at outcrop 20 are aa-lavas but approximately 5 m higher, a sequence of at least four non-brecciated lava flows is present. Non-brecciated flows have relatively smooth and horizontal contacts. The tops of flows are highly vesicular, whereas basal parts clearly contain fewer and smaller vesicles. These lava flows are also dark brownish-gray in color but contain smaller amphibole and pyroxene phenocrysts as well as small plagioclase crystals that delineate a crude flow structure.

Above the low viscosity lava flows, at an elevation of 6890', rounded quartzite and mica-gneiss boulders (up 1.2 m in diameter) are present on the surface. The boulders are scattered in a horizon that is a few tens of meters wide and over 300 m long, suggestive of an unexposed conglomerate bed. This may also simply be a result of Quaternary landslides or rock falls (Beaty, 1989).

At least two more lava-breccias (aa-flows) can be distinguished at the top of hill 6961 (outcrops 21 and 22). These flows are brownish-gray andesites with 10-15 percent typically 1-5 mm (but up to 1.5 cm) long pyroxene phenocrysts and abundant (15 percent) very small (< 0.5 mm) plagioclase crystals.

Other studied outcrops along the northeast side of Indian Hollow include conglomerates at outcrop 23, lava breccias at outcrop 24, and lava flows at outcrops 25 and 26. These rocks are similar to conglomerates of outcrop 14, breccias of 24 and lava flows of 21 and 22, respectively. However, all of them are found at elevations that are 24 m (80') higher than the corresponding outcrops. Taking into account variations in the attitude of bedding, observed dike offsets, shear zones at outcrop 2, and the high elevation of felsic andesitic breccias at outcrop 4, it seems probable that faults are present in the surroundings of hill 6848. Therefore, the brecciated pyroxene andesite lava flows found at outcrop 27 (at an elevation of 7480', Appendix 2) are not as much higher in stratigraphy as the elevation difference between outcrops 22 and 27 suggests.
12.5 Northwest fork

A thick sequence of breccias was also found along the northeast side of Indian Hollow. The two lowest outcrops are thickly bedded pumice-rich breccias (outcrop 29). The beds are 1-2 m thick and massive. Grading was not observed. Besides pumice, the deposits also contain some dense volcanic rock fragments and ca. 10 percent sedimentary rock fragments (including red sandstone of the Nugget Formation and limestone). The grain size is within the small pebble range. A thin section contains a few mudstone clasts, plagioclase-amphibole-clinopyroxene porphyritic volcanic rock fragments with microcrystalline groundmass, and a clast that appears to be either a mud chip that is armored with volcaniclastic detritus (mostly small plagioclase fragments) or altered accretionary lapilli. Pumice clasts contain plagioclase, green amphibole (hornblende), and biotite phenocrysts and small feldspar laths, many of which are probably sanidine as suggested by their low relief. The sample stains readily to yellow, when a sodium cobaltinitrite stain is applied, which also suggests a high K-content. The sample contains less than 5 percent matrix.

The polymictic compositions suggest a reworked origin for these beds. They are relatively well-sorted and the amount of older rock fragments is small.

These deposits are overlain by apparently polymictic, dense andesite clast-rich breccias which are covered by lichens. Another pumice breccia is found at an elevation approximately 15 m higher (outcrop 32). In this breccia, larger pumice clasts (Ø 2-12 cm) are supported by a tuffaceous matrix (sample IH 32, Table 7). The rock contains ca. 5 percent matrix in the sedimentary petrological sense of the word. The rock is composed of minor clay-sized material, sand-sized volcanic rock fragments, and some crystal fragments of plagioclase, amphibole and biotite. Sand-sized volcanic rock fragments are dense or sparsely vesicular. The clast composition is virtually monomictic. They contain 15-20 percent plagioclase, 6 percent amphibole and 4 percent biotite phenocrysts. Amphibole phenocrysts have an 80° extinction angle. Phenocrysts are as long as 2 mm, but range from 0.5 to 1 mm. The original glassy groundmass is strongly altered to zeolites and clay minerals; it contains abundant smaller feldspar microcrystals, many of which (30-40 percent?) have low relief and probably are sanidine. A high K-content is also evident from bright yellow staining of the groundmass. In spite the fact that the phenocryst content suggests an andesitic composition, the chemical composition should be different from most of the studied
deposits. The high pumice content suggests that the deposits were derived from a highly viscous magma that was fragmented in an explosive manner. The high viscosity of magma may reflect higher SiO₂-content. A chemical analysis of this pumice breccia might indicate a rhyo-dacitic composition.

Outcrops 33-39 are all coarse polymictic, clast-to matrix-supported breccias and correlate with the breccia-dominated sequence exposed at the northeast side of Indian Hollow. All of these outcrops are quite covered by lichens and thus only modest observations were possible.

13. Deposits of Silver Creek Canyon

The approximately 150 m thick sequence of Keetley Volcanics is well exposed at Silver Creek Canyon, 5.5 to 6.5 kilometers southwest of Wanship along the Interstate Highway 80 (Wanship Quadrangle, the western half of section 2, T.1 S. and R.4 E. and the southeast corner of section 34, T.1 N. and R.4. E., Appendix 3). Polymictic (samples SCC 1-1, 8-1, 15-1, Table 7) and clearly matrix-supported conglomerates make up an overwhelming part of the sequence. Several meter thick clast-supported conglomerate beds form a minor proportion. The thicknesses of matrix-supported conglomerates vary from a few meters to several tens of meters. Two beds that could have a pyroclastic mode of deposition were observed including one 2 m thick massive clast-supported volcanic conglomerate in the lower part of the sequence (outcrop 4, Appendix 3) and one 1.5 m thick volcanic pebbly sandstone (at outcrop 15, Appendix 3). The former is very homogeneous with abundant angular to subangular grayish-white pumiceous plagioclase-amphibole porphyritic andesite clasts. The latter shows crude inverse grading of pumiceous clasts.

Most of the conglomerates are laterally continuous sheetlike bodies. A few beds fill 10-15 m deep channels that are several tens of meters wide. The proportions of sedimentary rock and volcanic rock fragments varies. Some of the beds found in the lower part of the sequence consist entirely of red and light gray quartzose sandstone and limestone pebbles and cobbles. Such beds have a distinctive reddish-brown color and can be traced laterally over 70 m (outcrop 5 Appendix 3). One bed contains almost exclusively volcanic rock cobbles (matrix sample SCC 8 1, Table 7); the redeposited origin of this bed is evident from variations in the mineralogy and the texture (especially phenocryst size and crystallinity of the groundmass) of the volcanic rock fragments.
Most of the conglomerates contain both older sedimentary rock and volcanic rock fragments. Some of the volcanic rock fragments contain plagioclase phenocrysts as long as 1.5 cm. The matrix is clearly muddy and commonly calcareous. The clasts are are variably rounded. The sedimentary rock clasts are typically subrounded to well-rounded, whereas volcanic rock fragments are poorly rounded. The largest clasts are ca. 1.5 m in diameter. Bed thicknesses vary from a few meters to over 20 meters. Some of the apparently very thick beds have crude coarser horizons suggesting that they may really consist of several beds. The beds are generally massive; if inverse grading is present, it is restricted to a thin basal part that is less than 10 percent of the bed thickness. Two medium-coarse thin sandstone beds were found between matrix-supported conglomerate beds.

At the top of the sequence, the beds contain abundant angular cobbles and blocks of dark-colored pyroxene-amphibole porphyritic basaltic-andesite volcanics and red and light-gray quartz sandstone. Such beds contain quartz sandstone blocks as long as 5 m. The clasts are poorly rounded.

14. Summary of the studied deposits

The studied deposits are almost invariably andesitic, based on their mineralogy. Abundant microscopic low relief feldspar crystals in the groundmass and distinctive staining to yellow (after cobalt nitrate treatment) suggest a K-rich geochemistry. Probably most of the studied samples are trachyandesitic and those richer in mafic minerals may be shoshonitic based on their chemical compositions. Dacitic rocks are of minor importance. Lava flows, scoriaceous breccias and dikes have more mafic (basaltic andesite) compositions.

Summary stratigraphic sequences are given in Figure 48. The total thickness of the studied sequence at Jordanelle dam is over 450 m. Only a few of the exposures (21 outcrops, 9 roadcuts) represent the same stratigraphic levels. Many deposits show conclusive evidence of a reworked mass flow origin. For a great number of the beds, the origin could not be determined with certainty; they may be reworked mass flow deposits or pyroclastic flow deposits containing abundant accidental clasts. Fluvial deposits are minor. Lava flows are present only at stratigraphically high levels; some of them have reddish-brown surfaces indicating oxidation and proximity to the vent, but the source may have been a nearby parasitic vent. Debris flow deposits fill deep
channels and include well-rounded cobbles. Blocks of previously existing conglomerates or coarse sandstones were not observed; most likely rounded clasts were picked up by the debris flows as in recent debris flows described by (Scott, 1988). Also, breccia and minor sandstone deposits are similar to hyperconcentrated flow deposits (Smith, 1986; Scott, 1988; Smith and Lowe, 1991). Breccia beds are commonly inversely graded and less commonly inversely to normally graded. Some sandstone beds are normally graded beds. Some of the large blocks show radial in situ cooling-joints; transportation was not long enough to allow large blocks to cool before deposition. On the other hand, transportation was far enough to allow the grading patterns to be developed.
Figure 48. Illustration of the stratigraphy in the studied deposits at Jordanelle, Indian Hollow and Silver Creek Canyon.
Some of the shallow intrusions cut hydrothermally altered deposits. Others are themselves altered. Such observations are consistent with age data of Bromfield et al. (1972) and indicate that magmatism and hydrothermal alteration were overlapping events.

A few outcrops studied west of Francis and in the Provo River Valley further support the idea that individual breccias are not laterally continuous. The breccias at Provo River Valley include blocks with very large plagioclase phenocrysts. Such deposits were not found at other studied outcrops.

Thick bedded breccias of Indian Hollow are typically inversely graded. Many of them are polymictic. The coarse deposits found in Indian Hollow could be various coarse proximal breccias based on their grain-size distribution. Such breccias have good preservation potential only if capped by welded deposits. If capped, their recognition potential should be good (Cas and Wright, 1987, p. 482). However, gas escape structures, bomb sags or welding were not observed. At Indian Hollow, bedding is more characteristic of the internal stratification of very thick lag-fall units (Walker, 1985). Bed thicknesses are thinner than in typical proximal lag-breccias, and the beds are not interbedded with distinctive ignimbrites. The deposits might be explained, with difficulty, as lithic concentration zones or ground layers of proximal ignimbrites, but the associated lithologies do not support such an interpretation. The beds do not grade upward to matrix-supported and lithic-poor breccias (i.e., the upper part of layer 2b of Sparks, 1976). Other associated beds are not distinctive pumice flow deposits, but are instead probably sandy debris flow or hyperconcentrated flow deposits. Therefore, the breccia-dominated sequence forming the lowest 50 m of the studied sequence at Indian Hollow are most likely mostly reworked breccias. However, the breccias with a monomictic clast lithology could be co-ignimbrite or block-and-ash flow deposits (Cas and Wright, 1987, pages 482-483; Smith and Roobol, 1990). The overlying basaltic andesite flows and scoria breccias indicate at least a local change in the composition of the erupting magma. Change of composition also changed the pyroclastic fragmentation processes.

The sequence at Silver Creek Canyon along U.S.1-80 was almost completely deposited by cohesive debris flows. Clast-supported conglomerates are debris flow deposits in which cohesion probably was not the dominant clast-support mechanism (Nemec and Steel, 1984; Cas and Wright, 1987, pages 323-328; Smith and Lowe,
Thin sandstones represent hyperconcentrated flows associated with debris flows. Debris flows containing almost entirely older sedimentary rock clasts resulted from gravitational instability of the sediments covering nearby and already uplifted terrain. Debris flows representing mixtures of volcaniclastic material and rounded sedimentary rock clasts were formed from bulking of rain or melt water floods by fluvial sediments to form hyperconcentrated flows and, finally, debris flows (Scott, 1988; Smith and Lowe, 1991). Flows dominated by large angular blocks of dark-colored basaltic andesite lavas may be related to natural dams (Smith and Lowe, 1991). As in the Jordanelle and Indian Hollow sequences, evidence of more mafic (basaltic andesite) magmatism was found in the upper part of the sequence.

Some of the shallow intrusions cut hydrothermally altered deposits. Others are themselves altered. Such observations are consistent with age data of Bromfield et al. (1972) and indicate that magmatism and hydrothermal alteration were overlapping events.
15. Discussion

If the deposits are lacking the grading patterns and depositional sequences that are characteristic either of pyroclastic deposits or of reworked deposits such as the ignimbrite sequences of Sparks' (1976) or the transitional flow or lahar-run out facies of Scott (1988), the determination of the depositional mode strongly relies upon the distinction between polymictic and monomictic clast compositions. The presence of rounded boulders, cobbles and pebbles suggests deposition from mass flows that picked up such clasts from underlying fluvioglacial deposits. A reworked origin can be also postulated if deposits contain pieces of rocks that are columnarily jointed or altered prior to deposition. In situ cooled, radially jointed clasts, abundant gas escape structures and burned wood are characteristic of pyroclastic deposits but they have also been recorded in mass flow deposits.

Ascertaining of a polymictic composition was not easy in some of the studied rocks of the Keetley Volcanics. Different clasts lithologies can be defined based on variations in mineralogy and texture. However, the tentative field results had to be verified by petrographic studies of the clasts. These thin section studies were time consuming and set limits to the petrographic studies of the matrix samples. Different clast lithologies can be defined, based on variations in mineralogy and texture.

By studying the roundness of pebbles in the reworked deposits in the surroundings of Mt. St. Helens, Scott (1988) was able to indicate that the large proportions of these clasts were derived from underlying fluvioglacial sediments. Since a long transportation is needed to create a good rounding of sand, roundness is a poorly applicable criteria to distinguish such sand-sized bulked up grains. However, sand-sized grains are more easily eroded than coarser pebbles and bulked up clasts are likely more abundant in the sand-fraction. One might therefore, expect such deposits to have a distinctively polymictic matrix. Therefore, it is not surprising that many of the studied matrix samples from the Keetley Volcanics that were derived from the beds showing megascopic reworked evidence also are polymictic.

In contrast, in the pyroclastic deposits described by Smith and Roobol (1990) in the vicinity of Mt. Pelée, less than 10% of granules or smaller grains were lithic that is, derived from previously existing rocks.
Single eruptions of compositionally zoned or mixed magma may produce clasts with variable texture, mineralogy and chemical composition, and thus complicate the interpretation of the petrography. In spite of all the primary and secondary factors that may affect the texture of volcaniclasts, detailed matrix petrography of modern deposits might provide criteria applicable to define the depositional mode of the ancient deposits. However, the studies to which the Keetley Volcanics were compared are based mainly on megascopic studies of depositional structures.

Volcanic deposits and land forms are strongly dependent on eruption style. The eruption style in turn, is strongly dependent on SiO₂ and volatile content of the magma. The magma composition, the eruption style, clasts shapes and sizes, lithologies and large-scale sequences have close relationships which are also reflected in the studied deposits of the Keetley Volcanics.

Thick pumiceous deposits contain quartz phenocrysts and/or show an important potassium content. Eruptions of acid magma are commonly highly explosive and produce in addition to pumice, large amounts of sand-sized volcaniclasts. Thus, it is not surprising that the stratified pumiceous volcaniclastic sandstones of roadcuts 2 and 3 at Jordanelle (Appendix 1), although probably reworked, have a dacitic mineralogy. Also the association of basaltic-andesitic lava flows and scoria deposits result from at least a local change in the composition of erupted magma.

Thick breccia sequences comparable to the Keetley Volcanics are found on the flanks of stratovolcanoes and around intermediate to silicic-multivent centers (Cas and Wright, 1987, p. 427-431). Thick breccias are also formed in calderas associated with highly explosive eruptions. In addition to proximal pyroclastic breccias and ignimbrites, calderas are filled by intercalated caldera wall collapse breccias and fluvial and lacustrine reworked deposits (Lipman et al., 1976). Hydrothermal activity is considerable, and precipitates such as sulfide and sinter deposits are common. Collapse breccias are massive and characteristically very thick (tens to over 100 m thick). Vast volumes of pumiceous flow and fall deposits are susceptible to reworking processes. Acid volcaniclastics are associated with rhyolite domes and intrusions. Basalts or mixed pumice of acid and basic glass indicate bimodal magmatism. The resulting depositional sequence is complex. Although the thick breccias and conglomerates of the Keetley Volcanics rest on the top of the Peoa Tuff that represents volcaniclastic lake deposits (Woodfill, 1972; Bryant, 1990), the studied stratified andesitic to basaltic-andesitic
sequences do not otherwise support a caldera collapse origin for the Keetley Volcanics (Lipman et al., 1976; Cas and Wright, 1987).

Stratovolcanoes and intermediate-silicic multivent centers have variable eruption styles and fragmentation products. Larger volcanoes commonly have parasitic vents. Fall and flow deposits may be distributed around volcanoes quite asymmetrically. Domes and radial dikes are frequently found cutting volcaniclastic piles. The stratigraphic piles are also affected by caldera collapses, section failures (avalanches) and reworking mass flow and tractional processes. Volcaniclastics make up 70-80 percent of the successions on many slopes (Cas and Wright, 1987, p. 383). The distribution of different volcaniclastic deposits depends on position relative to the source. The core facies of Vessel and Davies (1981) includes intercalated intermediate to basic lavas, coarse to fine fallout tephra, and colluvial breccias caused by erosion and failure of steep flanks. In areas where erosion has deeply dissected volcanoes, "a vent facies can be identified by poorly bedded or massive pyroclastic breccias and collapse breccias in intrusive relation with the basement rocks" (Fisher and Schmincke, 1984, p. 358). In addition, fumarolic alteration may be extensive and one or more circular to elongated intrusions may be unroofed. Proximal facies (Vessel and Davies, 1981) around the source involves volcanic breccias and air-fall deposits. Breccias include block and ash flow deposits and colluvium breccias. Proximal pyroclastic flow deposits are commonly confined by erosional channels. With increasing distance from the vent, reworked pyroclastic deposits become prominent and the deposits may include epiclastic material. Avalanches, channeled to lobate debris flows, rivers and unconfined floods pile up volcaniclastic alluvial fans (Vessel and Davies, 1981; Fisher and Schmincke, 1984; Smith, 1991). Volcaniclastic alluvial fans may still be reached by pyroclastic and lava flows that commonly form lobate- or tongue-shaped deposits (medial facies, Vessel and Davies, 1981; intermediate facies, Fisher and Schmincke, 1984). Fans feed relatively low-relief plains commonly referred to as ring plains or volcaniclastic aprons (Smith, 1991) on which tuffaceous sediment is transported by sinuous or braided rivers or unconfined floods. Repeated aggradation and avulsion may lead into low relief fan-shaped accumulations of sand. These accumulations deviate from much larger and thicker volcaniclastic alluvial fans by lacking downslope decrease in grain size and by having very low fluvial gradients.

The studied deposits near Jordanelle dam included very thick more or less channeled polymictic debris and hyperconcentrated flows which relate to the medial facies of
Vessel and Davies (1981). For many depositional units observed near Jordanelle dam, a reworked (debris flow or hyperconcentrated flow) origin is probable. If the other alternative, a primary pyroclastic depositional mode, had been chosen, the over-all sequence should have a proximal character.

The studied deposits at Silver Creek canyon can be almost unequivocally attributed to debris flows or hyperconcentrated flows. Thus the studied sequence has characters of the medial facies of Vessel and Davies (1981).

The studied stratified breccia deposits at Indian Hollow are not confined by steep channels. In spite of the beds being laterally continuous, gently channeled beds are common and indicate some type of mass flowage. The mass flows must have been able to carry large blocks and they must have had high yield strengths. The breccia-dominated sequence of Indian Hollow is therefore interpreted to have been deposited by lobate- or tongue-shaped debris flows or minor pyroclastic flows. This would link them to the medial facies of Vessel and Davies (1981). Woodfill (1972) considered Indian Hollow as a source area for the "flows" of the Keetley Volcanics. Some of the lava flows found at Indian Hollow might have been derived from the radial dike swarm. It should be noted however, that to the east of Jordanelle some of the lava flows have reddish-brown oxidation surfaces (Outcrop 21, Charcoal Canyon, Appendix 2) indicating a nearby vent. Also dikes with a similar, more mafic composition are present (Roadcut 9, Appendix 2). Therefore, lava flows may have erupted from several smaller parasitic vents.

As noted by Woodfill (1972) the geomorphology of Indian Hollow has undoubtedly been affected by post-volcanic fluvial erosion. Its landscape is reminiscent of U-shaped valleys resulting from lateral blasts similar to the 18 May, 1980 eruption of Mt. St. Helens. The Indian Hollow plug (Woodfill, 1972; Bryant, 1990) is correctly situated to be a remnant of a horizontally-blasting cryptodome. Lateral-blast-like eruptions are also promoted by dilation around parallel dike swarms, although in the case of Indian Hollow, the dikes are oriented radially. Other factors that could have favored instability of the volcanic slopes at Indian Hollow include hydrothermal alteration around the plug at the base of the valley (Woodfill, 1992, personal communication) and massive lavas overlying a pyroclastic substratum (Siebert, 1984).

The studied breccia deposits seem to be similar to debris flow, hyperconcentrated flow and pyroclastic flow deposits that are found at some distance (kilometers to few
tens of kilometers) from the source on slopes of modern stratovolcanoes (Vessel and Davies, 1981; Scott, 1988). Some of the material may have been derived from local explosive to gravitational collapses of spines and domes related to shallow intrusions cutting the breccia. Basaltic-andesite flows and scoria breccias are at least partly derived from local parasitic vents. The position of basaltic-andesite lava flows at stratigraphically high levels may, however, indicate compositional zoning in a chamber. Indications of paleoflow were not present.

In spite of the fact that the various porphyritic plutons in the Wasatch mountains, 5 to 20 km west of the study area, have younger K-Ar and fission track ages from biotite (Crittenden et al., 1973; Bromfield et al., 1977), the intrusions indicate the most probable source area; these plutons may represent unroofed remnants of an ancient stratovolcano or intermediate-silicic multivent centers (vent facies, Fisher and Schmincke, 1984 p. 358, Cas and Wright, 1987, p. 382-394, 427-432). According to Bromfield et al. (1977), some of the age data are inconsistent with the observed geologic age relationships and some of the hornblende (K-Ar) ages from these intrusive rocks are older than similar ages from the Keetley Volcanics. The radiometric ages indicate the locking of the radiometric isotopes. In phaneritic intrusive rocks in which the cooling of the magma takes a long time, this is not necessarily the time of emplacement. On the other hand, the intrusive rocks may not represent the actual magma sources, but rather indicate the location of the magmatic center.

Gans et al. (1989) related the intrusions of the Uinta-Oquirrh mineral belt (Bromfield et al., 1977) and the Keetley Volcanics to an early Cenozoic syn-extensional tectono-magmatic event. The Uinta uplift was formed after a period of intensive overthrusting. If the Uinta uplift and normal faults are attributed to extensional tectonism following a compressional stage that overthickened the crust, then the volcanism apparently postdates the onset of the change in orogenic style, for the Uinta uplift already had supplied material to the Paleocene-Eocene Wasatch Formation (Bryant, 1991). The studied samples (ca. 100 thin sections) of the Keetley Volcanics also lack conclusive petrographic evidence of magma mixing. The model of Gans et al. (1989) emphasizing the spatial and temporal relations between magmatism and extension, may still be applicable to the Keetley Volcanics. The volcanics may have been derived from a larger compositionally zoned magma chamber formed when extension was well in progress. However, the abundant mafic enclaves and resorbed phenocrysts suggest that the magma differentiation has been a complex process. Geochemistry of the Keetley...
Volcanics and the intrusive rocks of the Uinta-Oquirrh mineral belt might provide interesting information about the temporal and spatial relations between regional Tertiary magmatism and tectonism.

16. Conclusions

The Keetley Volcanics include volcanic breccias, volcanic conglomerates, volcanic sandstones, lava flows and shallow intrusions. An overwhelming proportion of the Keetley Volcanics consists of volcanic breccias and conglomerates which were mostly derived from andesitic sources. Basaltic andesites and dacites were minor sources. The breccias vary from polymictic to monomictic. Minor source rocks of polymictic breccias and conglomerates include volcanioclastic rocks (volcanic breccias, volcanic sandstones and pumiceous tuffs) and epiclastic sedimentary rocks (red and light gray quartzose sandstone, and limestone).

Polymictic breccias and conglomerates most likely represent reworked volcanioclastic deposits that were emplaced at the site of deposition by debris flows, hyperconcentrated flows or streams. In many breccias that contain only andesitic clasts, the differentiation between the polymictic or monomictic composition is less evident. These breccias could have been deposited by pyroclastic transportation processes. In general, such deposits lack gas escape structures, ballistic bombs and welded or agglutinated clasts. The deposits do not show the facies association typical of lag-fall breccias or other co-ignimbrite breccias (Cas and Wright, 1987, p. 482). Therefore, such breccias were also most likely deposited by sedimentary mass flow processes. Studied volcanic sandstones are volumetrically minor and include both pyroclastic flows and reworked deposits. Lava flows have a more mafic (basaltic-andesitic) composition and they are commonly associated with scoria breccias of similar mineralogical composition.

The lava flows were fed from several, probably parasitic vents. Their appearance only at relatively high stratigraphic levels may indicate compositional zoning in a magma chamber. Volcanic breccias and sandstones of the different study areas of this project are similar to the deposits found on slopes of modern stratovolcanoes (proximal
and medial facies of Vessel and Davies, 1981, facies model for continental stratovolcanoes, Cas and Wright, 1987, p. 427-432). The source vent or vents in those models have been located kilometers to few tens of kilometers from the sites of deposition. The most probable sources for the Keetley Volcanics are the unroofed intrusives of the Uinta-Oquirrh mineral belt, although the shallow intrusions and dikes present at the study area may represent minor sources (vent facies, Fisher and Schmincke, 1984).
Appendix 1. Locations of the studied outcrops (OC 1-22) and roadcuts along the newly built highway 189 at Jordanelle, on the Heber quadrangle, in sections 31-33, T.2 S. and R.5 E, and sections 4-7 and 18, T.3 S. and R.5 E.
Appendix 2. Locations of the studied outcrops at Indian Hollow on the Kamas Quadrangle, in sections 11-15 of T.2S. and R.SE.
Appendix 3. Locations of the studied outcrops at Silver Creek Highway 5.5 to 6.5 kilometers southwest of Wanship along The Interstate Highway 80 on the Wanship Quadrangle, in the western half of section 2, T.1 S. and R.4 E.
Appendix 4. Locations of the studied outcrops at an old quarry near Francis (Francis Quadrangle, the northwest corner of section 32, T.2 S. and R.6 E.) and at roadcuts in the Northern Provo River Valley (Francis Quadrangle, the northern halves of sections 34 and 36, T.2 S. and R.5 E.).
References:


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