Late Miocene-Pliocene detachment faulting and Pliocene-Pleistocene Basin-and-Range extension inferred from dismemberedrift basins of the Salt Lake Formation

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Late Miocene-Pliocene Detachment Faulting and Pliocene-Recent Basin-and-Range Extension Inferred from Dismembered Rift Basins of the Salt Lake Formation, SE Idaho

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ABSTRACT

Geologic mapping, basin analysis, and tephra correlations in the Clifton and Malad City East 7.5-minute quadrangles in southeastern Idaho indicate that the modern Basin-and-Range topography is only a few million years old and was superimposed on unrelated rift basins associated with the ~12 to < 4 Ma Bannock detachment system. The Miocene-Pliocene Salt Lake Formation in the greater Cache Valley area was deposited during three sub-episodes of west-southwest extension on the Bannock detachment system.

Depositional systems within the Salt Lake Formation (>10.27 ± 0.07 Ma to < 5.1-4.4 Ma) evolved in response to the changing structural configuration of rift basins in the hanging wall of the Bannock detachment system. Early alluvial fans derived from the underlying Paleozoic rocks were replaced by broad, saline/alkaline lakes during the translation phase on the detachment fault. Later, as the hanging wall broke up internally, freshwater lakes and deltas occupied newly created NE-tilted half graben, and eventually filled them with braided streams. The Bannock detachment system collapsed the Cache-Pocatello Culmination of the dormant Sevier fold-and-thrust belt, much like the Sevier Desert Detachment collapsed the Sevier Culmination and the Wasatch Fault collapsed the Wasatch Culmination.

Deposition and tilting of the Salt Lake Formation ceased in the middle to late Pliocene (<2 Ma?) before uplift of the Clifton Horst along a new system of Basin-and-Range normal faults. This later episode of normal faulting uplifted and exposed metamorphic rocks in the footwall of the Bannock detachment system for the first time, and resulted in the deposition of < 200 m of Pliocene-Pleistocene(?) piedmont gravel and conglomerate deposits in angular unconformity on the exhumed bedrock of the horst and the adjacent, down-faulted Salt Lake Formation. External drainage may explain the sparse sedimentary record of Basin-and-Range faulting in this area.

The young age of both the large-magnitude extension (starting before 10.27 to and ending after 5.1 or 4.4 Ma) and of the modern Basin-and-Range topography (developing after 4.4 to or 5.1 Ma, probably 2 to 3 Ma and younger) in the Cache Valley region supports recent analyses that show both westward and eastward younging of extension from the central Basin-and-Range province. The synrift deposits in southeast Idaho illustrate a common evolutionary sequence in the Basin-and-Range province: 1) Deposition in broad basins primarily during early large-magnitude extension along listric low-angle normal faults, 2) formation of smaller half graben, reworking of older strata and continued deposition during break-up of the hanging wall of the low-angle normal fault along closely spaced normal faults, 3) and subsequent small-magnitude extension along steeper, widely spaced, N-striking range-front faults that dismember the earlier detachment system.
INTRODUCTION

The Salt Lake Formation, and the equivalent Starlight Formation, are exposed across large areas of Utah and southern Idaho. They have long been interpreted as the sedimentary record of Basin-and-Range extension (Fig. 1) (e.g., Williams, 1948; Slentz, 1955; Bryant et al., 1989). Poor exposure and inadequate geochronology have hampered past efforts to interpret the regional tectonic significance of these units. Recent tephra-stratigraphic analyses show that the Salt Lake Formation and its equivalents are mostly middle to late Miocene in age, and are younger toward the eastern and western margins of the Basin and Range province (Perkins et al., 1995; 1998; Stewart and Sarna-Wojcicki, 2000; Henry and Perkins, 2001). Although these data indicate the outward migration of block faulting from the present Utah-Nevada state line, in agreement with thermochronologic studies (Dumitr et al., 1997; Stockli, 2000), our understanding of this process is incomplete. The absence of basin-margin facies in many exposures of the Salt Lake Formation near modern range-fronts (e.g., Miller, 1991; Miller and Schneyer, 1994; Goessel et al., 1999; Stewart and Sarna-Wojcicki, 2000; Biek et al., 2001; Perkins, unpublished data), and a poor understanding of the structural-stratigraphic evolution of the basins filled with the Salt Lake Formation, indicate that more sophisticated analyses are needed to reconstruct the paleogeographic and the tectonic evolution of the Basin-and-Range province in the late Cenozoic. Understanding the age of initial normal faulting and the timing and nature of specific transitions in the structural style of extension will shed light on the ultimate causes of extension and allow conflicting tectonic models of Basin-and-Range extension to be assessed (e.g., Anders et al., 1989; Wernicke, 1992; Constenius, 1996; Humphreys and Hemphill-Haley, 1996; Stewart, 1998; Rodgers et al., in press). A strong spatial association between the Salt Lake Formation (and its equivalents) and highly extended terranes above detachment faults (see maps of Stewart and Carlson, 1978; Hintze, 1980; Mueller et al., 1999) suggest that low-angle normal faults may be genetically related to the Salt Lake Formation. The new data will also shed light on the interplay of sedimentation and tectonics within evolving rift basins.

Geologic mapping of the 7.5-minute Clifton and Malad City East quadrangles in southeast Idaho shows that facies patterns and provenance of the Miocene-Pliocene Salt Lake Formation are inconsistent with the modern geography of the area. We propose that the Salt Lake Formation was the primary basin-fill deposit of a Miocene-Pliocene rift basin that developed in the hanging wall of the Bannock detachment system. The basin was later segmented by younger, high-angle, Basin-and-Range normal faults. This paper examines the lithology, stratigraphy, and depositional environment of the Salt Lake Formation and younger Pliocene-Pleistocene (?) piedmont gravel deposits in order to determine the paleogeography of the Miocene-Pliocene rift-basin and the superimposed Basin-and-Range horst block.

The earliest studies of the Salt Lake Formation in Cache Valley were by Peale (1879), Gilbert (1890), and Mansfield (1920, 1927). Later studies in the region include Keller (1952, 1963), Adamson (1955), Adamson et al. (1955), Smith and Nash (1976), Danzl (1982, 1985), Sacks (1984), Sacks and Platt (1985), Smith (1997), Goessel (1999), Goessel et al. (1999), Oaks et al., (1999), Janecke and Evans (1999), and Crane (2000). Rodgers et al. (in press) characterized the stratigraphy equivalent units along the southern margin of the Eastern Snake River Plain and proposed that rift basins there not only roughly coincide with modern valleys but also young northeastern, in two waves of deformation. The major episode of extension occurred in advance of the Yellowstone Hotspot. Altogether the normal faults south of the Eastern Snake River Plain produced 15-20% extension according to Rodgers et al. (in press).

Closer to our study area, in the Portneuf Range of southeastern Idaho, Sacks and Platt (1985) suggested that the Salt Lake Formation filled accommodation space produced by the Valley Fault, a west-southwest-dipping listric normal fault that flattens at depth (Fig. 2). They also showed that the Salt Lake Formation was initially deposited in small, kilometer-wide sub-basins that later coalesced to form one large basin in the hanging wall of the Valley Fault. Janecke and Evans (1999) proposed that the Salt Lake Formation was deposited in rift basins, unlike the modern basins, above the Bannock detachment system. They interpreted the Valley Fault of Sacks and Platt (1985), and its along-strike continuations, as the breakaway of the Bannock detachment system. Megabreccia deposits, unroofing sequences, and provenance of conglomerates within the Salt Lake Formation in Cottonwood Valley (Sacks and Platt, 1985), the Lava Hot Springs area (Crane, 2000), the Oneida Narrows area (Danzl, 1982, 1985) and southern Cache Valley (Smith, 1997; Oaks et al., 1999) show the synextensional character of
Figure 1: Regional map showing the major active normal faults adjacent to the Eastern Snake River Plain (ESRP), the north-northwest-trending Cache-Pocatello Culmination (CPC), and selected other features. Note that the inferred Bannock detachment system essentially coincides with the Cache-Pocatello Culmination, and has the opposite sense of vergence as the partly coeval Raft River Metamorphic Core Complex. The aerially less extensive Wasatch Culmination (WC) began to collapse in latest Eocene time (Constenius, 1996). GSL = Great Salt Lake, WF = Wasatch Fault, YH = current position of the Yellowstone hotspot. Compiled from Schirmer (1988), Kuntz et al. (1992), Rodgers and Janecke (1992), Yonkee (1997), Stewart et al. (1998), and Janecke, unpublished subcrop map.
Figure 2: Simplified geologic map showing the distribution of the Salt Lake Formation around Cache Valley, Idaho and Utah, and the active Basin-and-Range normal faults. Some older normal faults are also shown (grayed). The Clifton Horst is the up-thrown block between the Deep Creek and Dayton-Oxford faults. Abbreviations are: CV = Cottonwood Valley, DCF = Deep Creek Fault, DCHG = Deep Creek half graben, DOF = Dayton-Oxford Fault, ECF = East Cache Fault (N = northern, C = central, S = southern segment), LM = Little Mountain, MC = Mink Creek, ON = Oneida Narrows, OP = Oxford Peak, QT = Quaternary-Tertiary sediment, RRP = Red Rock Pass, Tsl = Salt Lake Formation, WCF = West Cache fault zone (CM = Clarkston Mountain, JH = Junction Hills, W = Wellsville segment), WF = Wasatch Fault. Some buried faults (dotted) from Zoback (1983). Letters A-L refer to entries in Table 3. A-A’ and B-B’ are geologic cross sections in Figures 5 and 6, respectively. This map supercedes a compilation in Janecke and Evans (1999) and includes data from Goessel et al. (1999) and Oaks et al. (1999).
the Salt Lake Formation along the northeast and east margin of the area in figure 2. These studies also demonstrate a genetic link between the Salt Lake Formation and the Valley and East Cache faults.

Low-angle normal faults exposed within the Clifton Horst (Raymond, 1971; Mayer, 1979; Link 1982a, 1982b) have been correlated with the Valley Fault and interpreted as the uplifted, structurally lower portion of the Bannock detachment system (Janecke and Evans, 1999). This correlation is controversial, however, and this structural style may or may not persist laterally along strike to the Eastern Snake River Plain (c.f., Janecke and Evans, 1999; Crane, 2000; Kellogg et al., 1999; and Rodgers et al., in press, for different interpretations of the southern and northern ends of the extension system). In this paper we examine the Salt Lake Formation in order to evaluate these conflicting paleogeographic and tectonic models. Detailed interpretations of the structural evolution of the system of low-angle normal faults in this area are in Carney, 2002 and will be presented in a separate paper (Carney and Janecke, unpublished data).

This study expands on the initial stratigraphic and sedimentologic work of Janecke and Evans (1999) in the Deep Creek half-graben (Malad City East quadrangle and part of the adjacent Clifton quadrangle) (Figs. 2, and 3) and proposes a regional interpretation based on our new data and previous results from around the northern Cache Valley area. We describe the sedimentary record of an evolving detachment system (the Salt Lake Formation), the sparse sedimentary and erosional record of Pliocene to Recent Basin-and-Range extension, and the possible association between an earlier culmination in the Sevier fold-and-thrust belt and later, large-magnitude extension.

GEOLOGIC SETTING

Regional Geology

The Bannock Range in southeast Idaho is at the west edge of the Sevier fold-and-thrust belt of the North American Cordillera (Allmendinger, 1992), and lies at the northeastern margin of the Basin-and-Range province (Figs. 1, and 2). The study area is in the hanging wall of the west-dipping Paris-Willard thrust fault, which is part of the Idaho-Wyoming-Utah fold-and-thrust belt (Mansfield, 1927; Royse et al., 1975). The Paris-Willard thrust fault is exposed on the east margin of the Bear River Range approximately 45 km east of the study area, (Oriel and Platt, 1980; Dover, 1995) and was active from the Early to Late Cretaceous (Mansfield, 1927; Royse et al., 1975; Wiltshcko and Dorr, 1983; Oriel and Platt, 1980; Dover, 1995; Yonkee, 1997). The study area was transported eastward in the hanging wall of deeper thrusts that were active until the Paleocene (DeCelles, 1994).

A subcrop map of southeast Idaho shows that the Bannock detachment system is superimposed on a previously unnamed structural culmination of the Sevier fold-and-thrust belt that is bounded by the Malad Ramp on its west-southwest margin and the Logan Peak Syncline and Putnam Thrust on its east-northeast margin (Fig. 1) (Rodgers and Janecke, 1992; and this study). Culminations are regions within fold-and-thrust belts that are uplifted along ramps relative to adjacent areas. We here name this structure the Cache-Pocatello Culmination for its southern and northern extent. The Cache-Pocatello Culmination exposes Paleozoic carbonate rocks as old as the Middle Cambrian Blacksmith Limestone in its most uplifted core (but no Neoproterozoic rocks, in contrast to Rodgers and Janecke, 1992) and extends over 150 km north-south before ending beneath the Eastern Snake River Plain. In the study area the Cambrian Nounan Formation is the oldest unit unconformably overlain by Tertiary rocks. Older Middle Cambrian Blacksmith Formation underlies the unconformity to the east (Sacks and Platt, 1985). The Cache-Pocatello Culmination has an east-west extent of 30-40 km in its present, extended, configuration, and was 25-30 km wide prior to extension. The southern portion of the culmination was imaged on a reflection seismic line as a south-dipping homocline beneath the central part of Cache Valley (Fig. 12 of Evans and Oaks, 1996). Stratigraphic evidence for the erosion of 4-7 km of pre-Tertiary rocks from the culmination is briefly described below.

At least two episodes of extension followed Sevier contraction in the western United States (Wernicke, 1992; Stewart, 1998). The first episode is characterized by regionally extensive, low-angle normal faults that accommodated large magnitudes of general east-west extension. That episode was marked by the initial development of metamorphic core complexes and associated calc-alkaline magmatism in the western United States. The structural style of the proposed Bannock detachment system was similar to that of this first episode of extension.
Figure 3: Simplified geologic map of the Clifton and Malad City East 7.5 minute quadrangles. See Figure 2 for location. The contact between Cache Valley and Third Creek members may be angular in some places (see text).
Figure 4: Schematic stratigraphic columns and clast counts of the Salt Lake Formation across the Clifton Horst and in the Deep Creek half graben. Note that clast counts from east of the horst show an unroofing sequence of the source area(s) vertically upward within the Third Creek Member whereas clast counts from the Deep Creek half graben show an unroofing sequence vertically upward through the whole of the Salt Lake Formation. Maximum thicknesses are shown.
Table 1

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Member</th>
<th>Location (Township and Range /Lat and Long/UTM)</th>
<th>Ash Bed type*</th>
<th>Age Range or Age (Ma)**</th>
<th>Comments (stratigraphic position based on contacts and structure)</th>
</tr>
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<tr>
<td>jce156-98</td>
<td>Skyline</td>
<td>SE/4, SW/4, SE/4, Sec 8, T15S, R37E (42° 07' 41&quot; N, 112° 09' 48&quot; W) 46°46′500 N, 403′500 E</td>
<td>Twin Falls “d2&quot;</td>
<td>10.5-8.0</td>
<td>Stratigraphic position is likely high in section within Ts</td>
</tr>
<tr>
<td>jce187-98</td>
<td>Cache Valley (Tcv)</td>
<td>SE/4, NE/4, SE/4, Sec 31, T14S, R37E (42° 09' 34&quot; N, 112° 10' 24&quot; W) 46°7′200 N, 403′400 E</td>
<td>Twin Falls Wooden Shoe Butte?</td>
<td>10.13 ± 0.03</td>
<td>Stratigraphic position is likely relatively high within Tcv</td>
</tr>
<tr>
<td>jce178-98</td>
<td>Cache Valley</td>
<td>SE/4, NIVW, NE/4, Sec 19, T14S, R37E (42° 11' 51&quot; N, 112° 11' 22&quot; W) 46°7′200 N, 403′400 E</td>
<td>Twin Falls Wooden Shoe Butte?</td>
<td>10.13 ± 0.03</td>
<td>Stratigraphic position is high within Tcv</td>
</tr>
<tr>
<td>jce122-98</td>
<td>Cache Valley</td>
<td>SW/4, SE/4, SE/4, Sec 36, T13S, R36E (42° 14' 33&quot; N, 112° 11' 34&quot; W) 46°7′700 N, 401′700 E</td>
<td>Twin Falls Wooden Shoe Butte?</td>
<td>10.13 ± 0.03</td>
<td>Stratigraphic position is very high within Tcv</td>
</tr>
<tr>
<td>suj303-98</td>
<td>Cache Valley</td>
<td>SE/4, NW/4, SE/4, Sec 17, T14S, R37E (42° 12' 11&quot; N, 112° 09' 10&quot; W) 46°7′300 N, 404′000 E</td>
<td>Twin Falls Wooden Shoe Butte?</td>
<td>10.13 ± 0.03</td>
<td>Stratigraphic position is likely high within Tcv</td>
</tr>
<tr>
<td>jce202-99</td>
<td>Third Creek (Ttc)</td>
<td>SE/4, NW/4, SE/4, Sec 17, T14S, R37E (42° 12' 11&quot; N, 112° 09' 10&quot; W) 46°7′300 N, 404′000 E</td>
<td>Twin Falls “a”</td>
<td>10.5-8.0</td>
<td>Stratigraphic position is approximately mid-Ttc <em>this sample correlates with 39-97</em></td>
</tr>
<tr>
<td>jce181-98</td>
<td>Third Creek</td>
<td>SW/4, SW/4, SE/4, Sec 31, T13S, R37E (42° 12' 19&quot; N, 112° 09' 26&quot; W) 46°7′200 N, 404′200 E</td>
<td>Twin Falls “a”</td>
<td>10.5-8.0</td>
<td>Stratigraphic position is likely very low within Ttc</td>
</tr>
<tr>
<td>jce166-98</td>
<td>Third Creek</td>
<td>SE/4, NE/4, SE/4, Sec 9, T14S, R37E (42° 13' 02&quot; N, 112° 09' 48&quot; W) 46°7′600 N, 406′000 E</td>
<td>Twin Falls “a1”</td>
<td>-8.1 ± 0.2</td>
<td>Stratigraphic position is very high within Tcc (This age conflicts with age of sample jce209-99)</td>
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<tr>
<td>jce53-97</td>
<td>Third Creek</td>
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<td>Twin Falls “b” (?)</td>
<td>10.5-8.0</td>
<td>Stratigraphic position is unknown within Ttc</td>
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<td>jce50-97</td>
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<td>Twin Falls “b”</td>
<td>10.5-8.0</td>
<td>Stratigraphic position is unknown within Ttc</td>
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<td>jce204-99</td>
<td>Third Creek</td>
<td>NW/4, NE/4, SW/4, Sec 17, T14S, R37E (42° 12' 19&quot; N, 112° 09' 36&quot; W) 46°7′200 N, 404′200 E</td>
<td>Twin Falls?</td>
<td>8 ± 0.3</td>
<td>Stratigraphic position is approximately mid-Ttc</td>
</tr>
<tr>
<td>jce137-98</td>
<td>Third Creek</td>
<td>SW/4, NE/4, SE/4, Sec 20, T14S, R37E (42° 11' 22&quot; N, 112° 09' 00&quot; W) 46°7′100 N, 403′100 E</td>
<td>Twin Falls?</td>
<td>10.5-8.0</td>
<td>Stratigraphic position is low within Ttc</td>
</tr>
<tr>
<td>jce121-98</td>
<td>Third Creek</td>
<td>NW/4, SE/4, NE/4, Sec 28, T14S, R37E (42° 10' 51&quot; N, 112° 07' 53&quot; W) 46°7′200 N, 406′300 E</td>
<td>Twin Falls Inkom</td>
<td>-8.3 ± 0.5</td>
<td>Stratigraphic position is unknown within Ttc</td>
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<td>jce58-97</td>
<td>Third Creek</td>
<td>NE/4, NW/4, SW/4, Sec 14, T14S, R37E (42° 10' 34&quot; N, 112° 07' 30&quot; W) 46°7′200 N, 406′800 E</td>
<td>Twin Falls</td>
<td>10.5-8.0</td>
<td>Stratigraphic position is unknown within Ttc</td>
</tr>
<tr>
<td>jce209-99</td>
<td>Third Creek</td>
<td>NE/4, SW/4, NE/4, Sec 9, T14S, R37E (42° 13' 22&quot; N, 112° 10' 28&quot; W) 46°7′500 N, 406′000 E</td>
<td>Heise</td>
<td>7.5-4.0</td>
<td>Stratigraphic position is very high within Ttc</td>
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<tr>
<td>jce207-99</td>
<td>Third Creek</td>
<td>SE/4, SW/4, NW/4, Sec 21, T14S, R37E (42° 11' 35&quot; N, 112° 08' 42&quot; W) 46°7′800 N, 405′300 E</td>
<td>Heise Santee/Kilgore</td>
<td>5.1 ± 0.12</td>
<td>Stratigraphic position is likely high within Ttc</td>
</tr>
<tr>
<td>smc55</td>
<td>Third Creek</td>
<td>SW/4, SE/4, SE/4, Sec 34, T14S, R38E (42° 10' 04&quot; N, 112° 08' 42&quot; W) 46°7′800 N, 405′300 E</td>
<td>Twin Falls</td>
<td>10.5-8.0</td>
<td>East of Clifton Horst</td>
</tr>
<tr>
<td>smc33</td>
<td>Third Creek</td>
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<td>Twin Falls</td>
<td>10.5-8.0</td>
<td>Ash interbedded with lacustrine limestones within the Clifton Horst in hanging wall of the Clifton detachment fault</td>
</tr>
<tr>
<td>smc58</td>
<td>Third Creek</td>
<td>NE/4, NE/4, SW/4, Sec 10, T15S, R37E (42° 10' 04&quot; N, 112° 08' 42&quot; W) 46°7′800 N, 405′300 E</td>
<td>Twin Falls</td>
<td>10.5-8.0</td>
<td>Southwest side of Clifton Horst</td>
</tr>
</tbody>
</table>

Table 1 continued on next page
The second episode of extension is characterized by moderately to steeply dipping, planar to listric normal faults that define typify the present active Basin-and-Range province. These normal faults bound mountain ranges that trend generally north-south, and are regularly spaced about 30 km apart between the Colorado Plateau and the Sierra Nevada Range (e.g., Stewart, 1998). The Bannock and Malad ranges are bounded both on the east and west by north-striking, active Basin-and-Range normal faults (Figs. 2, 3). These faults offset and expose the Miocene-Pliocene Salt Lake Formation and Pliocene-Pleistocene (?) piedmont gravels, as well as low-angle normal faults of the Bannock detachment system.

Local Geology

The Bannock Range forms the northwest flank of Cache Valley. Cache Valley is a north-trending, Basin-and-Range extensional basin, approximately 110 km long and 35 km wide (Fig. 2). The valley extends through northern Utah and southern Idaho, and is characterized by an east-tilted half-graben in the south that changes to a west-tilted half-graben in the north (Evans and Oaks, 1996; Janecke and Evans, 1999). The master fault in the northern part of the valley is the north-striking, east-dipping Dayton-Oxford Fault along the east side of the Clifton Horst (Figs. 5, 6 and 7). The northeast margin of Cache Valley is defined by several older, presently inactive, northwest-striking, southwest-dipping normal faults (Fig. 2).

METHODS

The structural evolution and paleogeography of the southern Bannock Range is based on: 1) detailed geologic field mapping in the Clifton and Malad City East quadrangles at a scale of 1:24,000 and reconnaissance mapping and reanalysis of existing data sets in adjacent areas (Fig. 3); 2) photogeologic mapping on 1:16,000 scale aerial photographs; 3) stratigraphic and sedimentologic analysis of Cenozoic basin-fill deposits (Fig. 4); 4) age determinations based on chemical correlations of tephras within the Salt Lake Formation with tephras of known age (Table 1); 5) geologic cross sections (Figs. 5, and 6); and 6) argon geochronology. The Clifton and Malad City East quadrangles contain the Malad Range, Clifton Horst, Deep Creek half graben, and northwest edge of Cache Valley (Figs. 2, and 3). A regional cross section at the scale of 1:100,000 was also constructed across the entire extensional system based on map data from Wach (1967), Platt (1977), Oriel and Platt (1980), Link and LeFebre (1983), Danzl (1985), Janecke and Evans (1999), Janecke, unpublished mapping, and this study (Fig. 5). Clast counts from conglomerates of the Salt Lake Formation were used to determine their provenance. The stratigraphy of the Salt Lake Formation in the Clifton quadrangle is similar to that in the adjacent Malad City East quadrangle. Therefore, the nomenclature and subdivisions of the Salt Lake Formation used by Janecke and Evans (1999) were applied across the area and extended to the east side of the Clifton Horst. The stratigraphic thicknesses of units in figure...
Figure 5: Regional geologic cross section A-A' of northern Cache Valley, Idaho area. The structure beneath Cache Valley is uncertain. Surface geology was compiled from Wach (1967), Platt (1977), Oriel and Platt (1968, 1980), Link and LeFebre (1983), Danzl (1985), Janecke and Evans (1999), Janecke, unpublished mapping, and this study. Peterson and Oriel (1970) provided some subsurface control. Restoration indicated about 60% extension. See Figure 2 for location of this section and section B-B'. Approximate position of the Malad Ramp (Rodgers and Janecke, 1992) is shown.
Figure 6: Geologic cross section B-B’ across the Clifton Horst in the study area. See Figure 3 for location of this section. Relationships beneath the Clifton Fault are uncertain. The Clifton Fault is the main structure of the Bannock detachment system in this area. Probable angular unconformity between Cache Valley and Third Creek members of the Salt Lake Formation is not shown because its geometry is not well characterized.
were determined from cross sections of the most complete and least deformed parts of the study area. Twenty-one tephra samples collected from the Salt Lake Formation in the study area were chemically correlated with 13 tephras of known age (Table 1). Chemical compositions of 20 glass shards were determined for each sample on the electron microscope at the University of Utah (Table 1)(methods of Perkins et al., 1995, 1998).

OVERVIEW OF STRUCTURAL GEOLOGY

A full description of the structural geology of the area is beyond the scope of this paper and is the subject of a separate effort (Carney, 2002). Few critical relationships are described below because they support our interpretation that the original, broad depositional basin of the Cache Valley Member of the Salt Lake Formation was disrupted at least twice by younger episodes of normal faulting (Table 2). We separate normal faults in this area into sets of normal faults, and groups of related structures. Each fault set formed during a distinct episode of extension, or time period. We identify three main episodes of extension in the area. These are listed from oldest to youngest and are as follows:

Episode 1, Fault Set 1: The oldest set of normal faults generally strike northwest and north-northwest and dip southwest to west-southwest (e.g. Fig. 8) except where they have been folded and dip to the east-northeast. Two main subsets of normal faults occur in this group: subset a: moderate to low-angle normal faults with Tertiary or Paleozoic rocks in their footwalls, and subset b: low-angle normal faults with metamorphosed Neoproterozoic rocks of the Pocatello Formation in their footwalls (Fig. 6). Structural arguments suggest that the low-angle normal faults of subset b underlie the entire Cache Valley region, were active at a low angle, and probably initiated at a relatively low angle (Janecke and Evans, 1999; Carney, 2002). However, exposures of these faults are limited to the Clifton Horst and Little Mountain area (Figs. 2, 5). We interpret faults of subset a as normal faults that formed in the hanging wall of detachment faults of subset b (Fig. 6). Subsequent offset by younger Basin-and-Range faults and overlap by Quaternary deposits, plus the dominant north-northeast to northeast dip of the Salt Lake Formation show that faults in fault set 1 are the oldest in the study area. The Clifton Fault is the master detachment fault in the Bannock detachment system (Figs. 5, 6).

Episode 2, Fault Set 2: Cross faults with east, northeast, and southeast strikes are fairly common in the region, and probably formed between episodes 1 and 3, but their age is not well established (Table 2; Fig. 3). Some of these cross faults appear to offset older northwest-striking normal faults. Most are truncated by the modern range-front faults. Some cross faults may have experienced multiple episodes of slip (e.g., Kellogg et al., 1999). Further research is needed to clarify their age and significance. Some of the east-northeast-striking cross faults may have formed in a Yellowstone-influenced strain field due to subsidence along and toward the Eastern Snake River Plain (McQuarrie and Rodgers, 1998; Janecke et al., 2000; Rodgers et al., in press).

Episode 3, Fault Set 3: Cross-cutting relationships and the morphology of the range fronts show that the youngest normal faults in the area strike north and dip both east and west. We follow Janecke and Evans (1999) and refer to these normal faults as Basin-and-Range faults (sensu strictu) because they created the modern topography. Some active normal faults with other strikes (northwest, northeast, east) are also included in set 3 (Figs. 2, 3). These may be reactivated faults that formed during previous episodes of extension. The Basin-and-Range normal faults bound steep range fronts, and offset Pliocene-Pleistocene (?) piedmont gravel deposits (Fig. 7b and 7d).

The following observations support our grouping of normal faults into three geometrically and temporally distinct sets: 1) Most normal faults of the set 1 dip west-southwest to southwest unlike the Basin-and-Range faults (set 3), which exhibit both easterly and westerly dip directions; 2) Many of the normal faults of set 1 are presently gently dipping, and portions of some faults are subhorizontal and east-dipping, whereas Basin-and-Range normal faults (set 3) are steeper (Fig. 6); 3) The strikes of the older normal faults (set 1) are northwest to north-northwest, about 30° counterclockwise from the strike of the modern, Basin-and-Range faults (set 3). North and south of our study area, the two sets of normal faults of (sets 1 and 3) have similar strikes (Figs. 2, 3) (Goessel et al., 1999; Oaks et al., 1999; Kruger et al., this volume); 4) Map and cross sectional analysis shows that some of the west-southwest-dipping normal faults of fault set
between closely spaced hanging-wall normal faults and an underlying detachment surface is best developed in the northern half of the Clifton quadrangle (Fig. 3). In contrast, the north-striking normal faults of set 3 cut the low-angle normal faults of set 1.
Figure 7 (opposite page and this page): Photographs illustrating the geomorphology of the Clifton horst and the associated Pliocene-Pleistocene (?) piedmont gravel deposits.

A. Steep east face of Clifton horst. No piedmont gravel deposits are preserved on Oxford Ridge or in Cache Valley in this view. Note the topographic contrast between the east and west sides (Figs. 7b, 7c, and 7d) of the Clifton horst. DOF = Dayton-Oxford Fault, with balls on downthrown side.

B. Dissected piedmont surface on west side of the Clifton horst. The position of the Deep Creek Fault is shown. P = piedmont surface, QTg = erosional remnants of the piedmont deposits preserved within the Clifton horst.

C. Piedmont surface on west side of Clifton horst.

D. Faulted piedmont surface in hanging wall of the Deep Creek Fault west of the Clifton horst.

E. Example of Pliocene-Pleistocene (?) piedmont deposits within the Clifton horst. The conglomerate overlies tuffaceous rocks of the Salt Lake Formation in angular unconformity, yet contains clasts of Brigham Group (white, angular clasts) and Pocatello Formation (Zp, dark groundmass and large, elongate clast). See Figure 8 for location of this exposure. Such consolidated exposures are rare.

F. Example of a piedmont deposit rich in carbonate-clasts from the Deep Creek half graben, in the hanging wall of the Deep Creek Fault (QTg in Figure 3). Bedding attitudes parallel the piedmont surfaces in 7c and 7d.
STRATIGRAPHY

Strata in the study area consist of Neoproterozoic to lower Paleozoic rocks unconformably overlain by the Eocene Wasatch Formation, the Miocene-Pliocene Salt Lake Formation, and Pliocene-Pleistocene (?) piedmont gravels and younger Quaternary deposits. Lacustrine sediments of Late Pleistocene Lake Bonneville cover low-lying areas east of the Clifton Horst in Cache Valley and west of the Clifton Horst in Malad Valley (Fig. 3).

Pre-rift Stratigraphy

Pre-Tertiary rocks in the study area are mainly exposed in the Clifton Horst and west of the Deep Creek half graben. These rocks include the Neoproterozoic Pocatello Formation, the Neoproterozoic Cambrian Brigham Group, and Cambrian to Ordovician carbonates and shales. Silurian to Permian rocks are exposed west of the study area in the Samaria Mountains (Platt, 1977). The irregular distribution of the Neoproterozoic and Paleozoic sedimentary and volcanic rocks in the study area allowed specific source areas to be identified within the Cenozoic deposits.

The Paleocene-Eocene Wasatch Formation is a synorogenic deposit of the Sevier fold-and-thrust belt (Oaks and Runnells, 1992). It crops out in the study area in six isolated fault blocks on the northwestern margin of the Clifton Horst west of Oxford Ridge. It probably occupied large east(?)-trending paleovalleys (Fig. 3). Within the Clifton Horst it unconformably overlies the Ordovician strata in five different fault blocks and the Cambrian strata in one.

The Wasatch Formation and overlying Salt Lake Formation overlie lower Paleozoic rocks of roughly the same stratigraphic level from the southern Portneuf Range (Sacks and Platt, 1985; Oriel and Platt, 1980) westward to Malad Valley. West of Malad Valley and also east of Cache Valley, in the core of the Logan Peak Syncline in the Bear River Range, Pennsylvanian and Permian rocks are preserved beneath the basal Tertiary unconformity (Platt, 1977; Rodgers and Janecke, 1992). These relationships indicate that 4-7 km of middle and upper Paleozoic and Mesozoic (?) rocks were eroded from the area between...
Malad Valley and eastern Cache Valley prior to deposition of the Wasatch Formation in Paleogene time (Rodgers and Janecke, 1992). This uplifted and eroded area is the Cache-Pocatello Culmination. The Sevier Culmination, which collapsed to produce the Sevier Desert detachment fault, exhibits a comparable structural relief along its western margin (DeCelles et al., 1995).

**Cenozoic Synrift Deposits**

Two distinct syntectonic sedimentary units younger than the Wasatch Formation are present in the area around the Clifton Horst: (1) Mio-Pliocene Salt Lake Formation and (2) Pliocene-Pleistocene (?) piedmont gravel deposits. Beds of the Salt Lake Formation comprise the bulk of the basin-fill deposits on either side and within the horst block. These deposits are at least 750 m to over 2 km thick (Fig. 4). Pliocene-Pleistocene (?) gravel and conglomerate deposits are a younger, newly identified unit within the Clifton Horst and along its western flank. We informally refer to these deposits as piedmont gravels because of their position adjacent to the Clifton Horst. Correlative deposits have not been identified in the fault block directly east of the Clifton Horst, but erosional remnants of elevated alluvial-fan deposits in northernmost Cache Valley (4 km southwest and 3.5 km southeast of Red Rock Pass [Fig. 2]) may be related to this the piedmont gravels. The thickness of the Pliocene-Pleistocene (?) piedmont gravel is variable (0 to >200 m thick) because the unit filled pre-existing erosional topography.

**Skyline Member**

The stratigraphically lowest member of the Salt Lake Formation, the Skyline Member, is restricted to the western margin of the Deep Creek half graben, and either pinches out eastward or occupied an early half graben within the western half of the younger Deep Creek half graben. The Skyline Member consists mostly of poorly sorted pebble to cobble conglomerate with tuffaceous, calcareous, and sandy matrix with interbeds of lacustrine limestones and variable amounts of tephra. The Skyline Member is up to 370 meters thick with most beds > 30 cm thick (Janecke and Evans, 1999). Clasts are composed mostly of lower Paleozoic carbonates with some chert and quartzite (Fig. 4). Conglomerates of the Skyline Member are alluvial-fan deposits (Janecke and Evans, 1999).

**Cache Valley Member**

The Cache Valley Member overlies the Skyline Member. It is exposed primarily on the west side the Deep Creek half graben and on the west side the Oxford Ridge in the hanging wall of a folded low-angle normal fault (Fig. 3). It is at least 335 m thick in the Pocket Basin on the west edge of the Clifton Horst (Fig. 4). The Cache Valley Member is composed mostly of reworked zeolitized and tuffaceous mudstone and siltstone, along with some limestone, silicified limestone, shale, sandstone, primary ash-fall tuffs and tephra, and rare pebble conglomerates (Fig. 4). Bedding in the unit ranges from thick, massive, and fractured to thin and shaley. Tephra beds are thin and are generally very light gray to white. The unit as a whole is generally light green to tan to white, with green being the most common. The green coloring is attributed to the zeolite mineral clinoptilolite. The zeolites are unique to the Cache Valley Member, and are only found in the Third Creek Member within clasts recycled from the Cache Valley Member.

Massive and parallel bedding characterize tuffaceous deposits of the Cache Valley Member in the study area. Sedimentary structures are uncommon but include current ripples, fining-upward sequences, and normal graded bedding. Nearly identical rocks in the Oneida Narrows area, about 20 km northeast of the Clifton Horst (Fig. 2), contain fining-upward sequences, parallel bedding and rare ripple marks (Danzl, 1985). Like Danzl (1985), we interpret these deposits as having formed in an open lacustrine environment. On the east side of Cache Valley, between Mink Creek and the Cub River, flute casts within the Cache Valley Member suggest deposition by turbidity currents in a deeper portion of the lake.

**Synrift Deposits – Main Episode**

The Salt Lake Formation in the Deep Creek half-graben consists of four members (Fig. 4)(Janecke and Evans, 1999). These distinctive members have now been identified across the entire study area, in the Clifton Horst and in a fault block along the western margin of Cache Valley (Fig. 3). Members of the Salt Lake Formation include the basal Skyline Member, interpreted as an alluvial-fan deposit; the Cache Valley Member, interpreted as a lacustrine deposit; the Third Creek Member, interpreted as a near-shore lacustrine, fluvial, and deltaic deposits; and the overlying New Canyon Member, interpreted as a fluvial deposit (Janecke and Evans, 1999)(Figs. 3 and 4). Altogether >2 km of Salt Lake Formation is are preserved in this area (Fig. 4).
The presence of limestone beds with algal lamina-
tions at the base and top of the Cache Valley Member
in the Deep Creek half graben (Fig. 4) indicates that
some portions of this unit were deposited in shallow
water. Open lacustrine conditions likely characterized
the tuffaceous middle portion of the member. The
notable scarcity of coarse detritus in the Cache Valley
Member indicates deposition in a position distant from
the uplifted margins of the basin. Rare pebbly con-
glomerate beds near the top of the Cache Valley Mem-
ber in the Deep Creek basin provide the only evidence
for an emerging highland in the vicinity. We interpret
the Valley Fault (Fig. 2) as the eastern margin of the
basin at this time because closer normal faults like the
Clifton Cemetery Fault appear to have bounded the
basin during deposition of overlying members, but
direct evidence for the Valley Fault during deposition
of the Cache Valley Member is lacking. Overall, the
Cache Valley Member is dominated by fine- to
medium-grained ash that fell into an open lake with
saline/alkaline conditions in deeper water (see below).
Lake level fluctuated, and water depth was greatest in
during deposition of the middle of the member.

Third Creek Member
The Third Creek Member overlies and interfingers
with the Cache Valley Member. This member is
exposed on the east side of the Clifton Horst in the
hanging wall of the Dayton-Oxford Fault and west of
the horst in the Deep Creek basin (Fig. 3). It is >1.1
km thick east of the Clifton Horst and at least 1 km
thick just west of the horst (Fig. 4). The Third Creek
Member is also exposed in the Clifton Horst in the
south-central section of the Clifton quadrangle just
west of Oxford Ridge (Fig. 3). It is composed mostly
of medium- to coarse-grained tuffaceous sandstone,
clast- and groundmass-supported conglomerate, and
poorly consolidated white to silvery-gray tephra.
Tephra beds vary from <20 cm to > 2 m, and individ-
ual conglomerate beds range from <50 cm to > 4 m
thick. Groundmass of the conglomerate is variably
tuffaceous, sandy, and calcareous. This unit also con-
tains interbedded, slightly tuffaceous, and locally
oolitic limestones with algal structures, shell frag-
ments, ostracod and gastropod molds, and chert nod-
ules. Clasts of the conglomerate are generally sub-
angular to well-rounded pebbles, and a few beds
contain cobbles (Fig. 9a).

Sedimentary structures in the Third Creek Member
suggest that it is a near-shore lacustrine and partly
fluvial deposit (Fig. 9a). An east-tilted coarse-grained
Gilbert-type delta forms the top of the Third Creek
Member near the eastern margin of the Deep Creek
basin (point A on Fig. 3) with northeast-dipping
topset beds and west-dipping foreset beds that indi-
cate transport from east to west (Janecke and Evans,
1999). Limestone beds with ooids, gastropod molds,
and shell fragments, and poorly consolidated tephra
beds are also present near the top and bottom of the
Third Creek Member in the study area. The middle of
the unit has conglomerate beds with well-sorted,
well-rounded clasts, and interbedded tuffaceous
sandstones with trough-shaped scours. The Gilbert-
type delta, freshwater limestone beds, and poorly
consolidated tephra beds suggest that the top and
bottom parts of the Third Creek Member are near-
shore deposits. Conglomerate beds with sandstone
interbeds indicate that the middle of the member is a
fluvial deposit. The Third Creek Member marks a
transition from a saline/alkaline lake environment of
the Cache Valley Member to a shallow freshwater
lake and river system.

Gilbert-type deltas form close to steep mountain
fronts with a large supply of sediment (Milligan and
Chan, 1998). A fault-bounded highland probably
was close to the Deep Creek half graben during
deposition of the Third Creek Member (Fig. 10b).

New Canyon Member
The New Canyon Member is exposed near Second
Creek in the northeast corner the Deep Creek half
graben and in two fault blocks on the east side of the
Clifton Horst (Fig. 3). The unit is poorly consolidated
in the study area. However, excellent consolidated
exposures of this member west of Oxford Peak, north-
west of the map area, show that the member there is a
parallel-bedded pebble to cobble conglomerate. The
conglomerates are clast-supported, and clasts are uni-
formly well-rounded with no fining- or coarsening-
upward sequences (Fig. 9c, 9d, 9e). Clasts consist
mostly of Brigham Group quartzites, Paleozoic car-
bonates and chert, and some recycled clasts from the
Cache Valley Member of the Salt Lake Formation
(Figs. 4, 11). The New Canyon Member is distin-
guished from the Third Creek Member by its lack of
tephras and freshwater limestones, and its distinct
brown to red color. The New Canyon Member is about
250 m thick east of the Clifton Horst (Fig. 4). Immedi-
ately north of the study area, more than 1.0 km may be
preserved west of Oxford Peak (Janecke and Evans,
1999). The New Canyon Member is interpreted to be a
fluvial deposit due to the presence of well-rounded
and well-sorted, equant to subequant clasts, its bed-
forms, and its persistence vertically and laterally.
Figure 9: Photographs of important conglomeratic units in the Salt Lake Formation.

A. Sub-rounded to rounded clasts of predominantly Brigham Group quartzites in conglomerate beds of the uppermost Third Creek Member of the Salt Lake Formation, west of Clifton Horst. Note the rather good sorting of each bed, and alternation between smaller and larger clasts in adjacent beds.

B. Photograph of angular, zeolitized clasts of the Cache Valley Member in a conglomerate of the Third Creek Member of the Salt Lake Formation. Recycled clasts like these are common in the Third Creek Member.

C. East-dipping beds of the New Canyon Member north of the study area in the hanging wall of the Deep Creek Fault. View to the south.

D. Detail of the New Canyon Member of the Salt Lake Formation in C. Note parallel and low-angle bedding and overall roundness of clasts derived from the Neoproterozoic Mutual Formation (Brigham Group). Although the Neoproterozoic Pocatello Formation is exposed along the crest of Oxford Ridge over a distance of about 28 km, its clasts are not found in the Salt Lake Formation in the Deep Creek half graben.

E. Cobble and boulder bed within the New Canyon Member of the Salt Lake Formation east of Weston Peak and the Clifton Horst.
Clast Counts in the Salt Lake Formation

Clast counts of conglomerates in the Salt Lake Formation, performed at the outcrop, were obtained for 2 locations in the New Canyon Member, 30 locations in the Third Creek Member, and 7 locations in the Skyline Member (Figs. 4, 11). Fifty to one hundred clasts >.5 cm were counted within a 1 m² area at each location. None of the conglomerates in the Salt Lake Formation contain clasts of the Neoproterozoic Pocatello Formation, a bedrock unit that crops out extensively in the Clifton Horst in the footwall of the Clifton detachment fault.

Clast lithologies in the Salt Lake Formation as a whole change upsection, and provide evidence for an unroofing sequence in the source area(s) (Figs. 4, 11). Lower Paleozoic carbonate clasts comprise an average of 99% of the basal Skyline Member, 41% of the Third Creek Member and 5% of the New Canyon Member. Neoproterozoic to Cambrian Brigham Group quartzite clasts are absent in the basal Skyline Member, but comprise an average of 32% of total clasts in the Third Creek Member and dominate the New Canyon Member with an average of 92% of total clasts. The composition of the Third Creek Member is extremely variable and probably records input from a variety of sources. Recycled tuffaceous clasts that resemble the Cache Valley Member, but are likely from tuffaceous beds in the Skyline Member, are nearly absent from the Skyline Member, and average only 1%. Clasts of recycled Cache Valley Member average 26% of total clasts in the Third Creek Member and dominate the New Canyon Member, but make up only 3% of total clasts in the New Canyon Member (Fig. 4).

In addition to the unroofing sequence documented in the Salt Lake Formation as a whole, there is evidence for unroofing of the source areas during the deposition of the Third Creek Member alone (Fig. 4). Clasts from locations 2 and 9 in the Third Creek Member show an upsection increase from 10% to 58% of clasts of Brigham Group quartzites and a decrease upsection from 90% to 31% of clasts of lower Paleozoic rocks (Figs. 4). Clasts from locations 10, 11 and 12 also show an increase upsection of the percentage of clasts of the Brigham Group quartzites from 11% to 34%, whereas clasts of Paleozoic carbonates decrease from 59% to 49%.

Diagenesis of The Salt Lake Formation and Its Paleoenvironmental Significance

Zeolites within the Salt Lake Formation may have formed in saline/alkaline lakes. Tuffaceous rocks of the Cache Valley Member of the Salt Lake Formation are altered to clay and zeolites in most locations. These rocks are greenish, off-white, yellow, and tan, and form partially indurated outcrops with a pervasive, randomly oriented, blocky fracture pattern. This is in marked contrast to silvery, gray, and white unaltered tuffs and tuffaceous sedimentary rocks of the Third Creek Member, which form soft unconsolidated to poorly consolidated exposures. Limestone beds within the Cache Valley Member are silicified to varying degrees. Weak silicification produced scattered chert nodules, whereas strong alteration resulted in the complete replacement of calcite by silica. The alteration of ash and limestone is so pervasive that the presence or absence of altered rock was one criterion used in the field to distinguish between the Cache Valley Member and younger members of the Salt Lake Formation.

Limited X-ray-diffraction analyses of materials from the study area show that green tuffaceous zones within the Cache Valley Member contain clinoptilolite, a Na-, K-, and Ca-bearing zeolite. Zeolites are hydrated aluminosilicates of the alkalies and alkaline earth minerals (Surdam, 1979).

More detailed analyses of lithologically identical rocks in the Oneida Narrows area (Danzl, 1985) and other locations (Table 3) show that clinoptilolite is the main zeolite species present within tuffaceous rocks of the Salt Lake Formation. Petrographic and XRD analysis of the diagenetic assemblages in the Oneida Narrows area showed that authigenic minerals develop in two separate alteration sequences: 1) hydration of glass and later partial crystallization formed authigenic silica and montmorillonite, and 2) hydration of glass, recrystallization of hydrated glass to authigenic silica and clinoptilolite, followed by total re-crystallization to clinoptilolite (Danzl, 1985). We assume that the same two processes operated throughout the depositional basin of the Salt Lake Formation.

Danzl (1982; 1985) interpreted the diagenesis as a secondary alteration product resulting from the percolation of groundwater, but also noted that significant changes in alteration assemblages coincide with vertical changes in the depositional environment of the parent rock. Tuffaceous beds in the Cache Valley Member are altered to clinoptilolite, whereas otherwise identical tuffaceous beds in the overlying Mink Creek Conglomerate (equivalent of the Third Creek Member?) lack clinoptilolite (Danzl, 1985). The Cache Valley Member was deposited in an open-lacustrine...
Table 3
LOCATIONS AND ATTRIBUTES OF ZEOLITE OCCURRENCES IN THE SALT LAKE FORMATION IN THE CACHE VALLEY AREA. SEE FIGURE 2 FOR LOCATIONS OF SAMPLES.

<table>
<thead>
<tr>
<th>Location</th>
<th>Quadrangle</th>
<th>Thickness of zeolite-bearing zone</th>
<th>Locations of unaltered tuffs relative to the zeolite-bearing zones</th>
<th>Source of data</th>
<th>Minerals, if known, and method of detection</th>
</tr>
</thead>
<tbody>
<tr>
<td>A. Deep Creek half graben</td>
<td>Malad City East, Idaho</td>
<td>Thick, about 400-600 m</td>
<td>Rare unaltered tephras interfinger with the zeolite-bearing beds near the top of the section</td>
<td>J. C. Evans and Janecke, unpublished data</td>
<td></td>
</tr>
<tr>
<td>B. Pocket Basin</td>
<td>Clifton, Idaho</td>
<td>381 m</td>
<td>Pervasive alteration</td>
<td>This study</td>
<td></td>
</tr>
<tr>
<td>C. Clifton Horst (excluding Pocket basin)</td>
<td>Clifton, Idaho</td>
<td>Variable, at least tens of m</td>
<td>Interfingering of zeolite-bearing beds with algal limestones and thick unaltered tephra</td>
<td>This study</td>
<td>Clinoptilolite, XRD</td>
</tr>
<tr>
<td>D. Gravel pit east of Clifton Horst at Five Mile Canyon</td>
<td>Weston Canyon, Idaho</td>
<td>Thin, &gt; 10 m</td>
<td>Relationship to tephras and limestone-bearing beds is uncertain</td>
<td>Janecke, unpublished mapping</td>
<td></td>
</tr>
<tr>
<td>E. Northern end of Oxford Ridge and environs Idaho</td>
<td>Scattered localities in Oxford and Malad Summit, Idaho</td>
<td>unknown</td>
<td>No data</td>
<td>Link, 1982a</td>
<td></td>
</tr>
<tr>
<td>F. Cottonwood Valley Treasureton, Idaho</td>
<td>Cottonwood Peak, Swan Lake, Medium thickness?</td>
<td>Mass lumata</td>
<td>Zeolites in two zones that alternate with unaltered tuffaceous beds at different scales and interfinger with lacustrine limestone beds (bed-by-bed and also groups of beds)</td>
<td>Sacks and Platt (1985); Sacks (1984)</td>
<td>Clinoptilolite, visual identification of concretions (p. 40 in Sacks and Platt, 1985)</td>
</tr>
<tr>
<td>G. Oneida Narrows</td>
<td>Riverdale, Idaho</td>
<td>380 m</td>
<td>Extensive alteration, especially at deeper stratigraphic levels, but degree of alteration is variable on a bed-by-bed level. Zeolites are restricted to the Cache Valley Member of the Salt Lake Formation</td>
<td>Danzl (1985)</td>
<td>Clinoptilolite, authigenic silica, montmorillonite, calcite, illite: XRD, petrography, SEM</td>
</tr>
<tr>
<td>H. Amoco Lynn Reese well</td>
<td>Smithfield, Utah</td>
<td>36 m</td>
<td>Unaltered tuffs underlie and overlie the zeolite-bearing bed</td>
<td>Brummer (1991)</td>
<td>Clinoptilolite, XRD</td>
</tr>
<tr>
<td>I. Junction Hills Clarkston, Utah</td>
<td>Cutler Dam, Idaho</td>
<td>98 m</td>
<td>Long Divide Zeolite subunit of the Cache Valley Member of the Salt Lake Formation. Unaltered tuff beds both interfinger with and underlie two zones of numerous zeolites</td>
<td>Goessel (1999); Oaks (2000)</td>
<td></td>
</tr>
<tr>
<td>J. Steele Canyon area Note: This is North Canyon of Adamson et al. (1955)</td>
<td>Henderson Creek, Idaho, and northernmost Clarkston Mountain, Utah</td>
<td>No zeolites</td>
<td>No data</td>
<td>Janecke; Biek; and Perkins, unpublished reconnaissance</td>
<td></td>
</tr>
<tr>
<td>K. Newton Mountain</td>
<td>Trenton, Utah</td>
<td>Unknown</td>
<td>No data</td>
<td>J. Boetting and P. Kolesar, (1999), unpublished data</td>
<td>Clinoptilolite, XRD</td>
</tr>
</tbody>
</table>
setting, whereas the overlying Mink Creek Conglomerate was deposited by as subaerial braided-streams (Danzl, 1985).

Two competing models may explain the presence of zeolites in the Cache Valley Member. The zeolites may have formed shortly after deposition of the tuffaceous deposits due to saline/alkaline lake conditions within their depositional basin (e.g. Surdam, 1979). Horizontal zonation of alteration assemblages should develop in this setting, and alteration would begin soon after deposition. This model has been applied to many occurrences of zeolite within tuffaceous lacustrine deposits in the Basin and Range province (Surdam, 1979; Sheppard, 1991, 1994). A second model, favored by Danzl (1985), suggests that downward percolating groundwater produces a vertical zonation of alteration minerals after deposition of some thickness of tuffaceous sedimentary rocks. The diagenesis would occur after deposition, and would be characterized by hydrated glass and montmorillonite at shallow levels and clinoptilolite at deeper levels.

For the Salt Lake Formation, several lines of evidence are more consistent with formation of the zeolites in a saline/alkaline lake than with post-depositional alteration by percolating groundwater:

1) Altered clasts of tuffaceous, zeolite-bearing sediment from the Cache Valley Member are common as clasts within conglomerates of the overlying unaltered Third Creek Member (Figs. 4, 9b, 11). Alteration of the Cache Valley Member must have occurred prior to deposition of the Third Creek Member because some of the clasts reworked from the Cache Valley Member are angular and preserve fractures that predate final deposition of the clasts (Fig. 9b). Unaltered equivalents of these siltstone and mudstone units are too poorly indurated to form coherent clasts, to survive appreciable transport, or to sustain or preserve fractures. XRD analyses confirm that the fractured clasts in Third Creek Member contain zeolite but the matrix does not;

2) The degree of alteration varies within the Cache Valley Member from pervasive to slight. Unaltered or weakly altered ashes alternate vertically with pervasively altered ashes in many localities. Unaltered ashes both underlie and overlie zeolite-bearing zones (Table 3). Fluctuating conditions within a saline/alkaline lake, with periodic freshening due to climatic or tectonic oscillations, could produce such interfingering relationships naturally. Downward groundwater percolation in a rapidly subsiding basin should result in a fairly steady, downward increase in the degree of alteration (Surdam, 1979; Danzl, 1985);

3) The presence of zeolites varies laterally across the study area at roughly the same stratigraphic level. The thickest and most pervasive zones of zeolite alteration coincide with a northeast-trending zone that stretches from the western Deep Creek half-graben to the area around Oneida Narrows (Table 3). South of the Deep Creek area, zeolite-bearing zones interfinger with unaltered ash (point L on Fig. 2), and even farther south, zeolites are uncommon or absent (point J on Fig. 2). Zeolites reappear farther south at sites K, H and I (Table 3 and Fig. 2);

4) The available geochronology suggests that the only two dated zones of zeolite, in the Deep Creek half graben, and in the Junction Hills (Oaks, 2000) are essentially the same age. In the Junction Hills the Long Divide zeolite subunit of the Cache Valley Member was deposited between about 10.3 and 9.2 Ma (Oaks, 2000). The Cache Valley Member in the Deep Creek half graben is older than 10.27 ± 0.07 Ma at its base and about 10.13 ± 0.3 Ma at its top (this study) (Table 1). A common tectonic or climatic cause probably explains the synchronous development of saline/alkaline lakes in regions that are now 35 km apart (Fig. 10a);

5) Zeolites appear to be more pervasive in tuffaceous deposits that formed in open-lacustrine conditions. Tuffaceous deposits that formed in marginal-lacustrine, fluvial, or subaerial settings contain freshwater molusks (Yen, 1947 and S. Good, written comm., 1998) and are less altered or contain no zeolites.

For these reasons we favor the saline/alkaline lake model for the development of zeolites in the Salt Lake Formation, but acknowledge that other models cannot be ruled out with the current data set.

Although ours is the first interpretation of saline/alkaline lakes within the Salt Lake Formation, this paleoenvironmental model is consistent with paleoclimatic evidence for arid climates in northern Utah from 10 Ma to the present (Davis and Moutoux, 1998). In addition, one of two pollen records from wells in the Great Salt Lake is dominated by the pollen of Sarcobatus vermiculitus (black greasewood) during the time period when zeolites are particularly abundant in the Cache Valley area (~10.5 to >9.6 ± 0.2 Ma; Table 3) (Davis and Moutoux, 1998). Greasewood is well adapted to ecological niches along the margins of saline/alkaline lakes (Dr. Neil West, Utah State University, oral comm., 2001). Sagebrush (Artemisia) replaced greasewood in younger parts of the core at a time when less altered tuffaceous deposit dominate in the Cache Valley area. If our interpretation were correct, it would indicate that depositional environments
within the Salt Lake Formation changed from closed hydrographic basins during deposition of the Cache Valley Member to more open systems later. Additional study is needed to test our hypothesis.

**Age of the Salt Lake Formation**

Estimation of the age of the Salt Lake Formation is based on $^{40}\text{Ar}/^{39}\text{Ar}$ dating of an ash-flow tuff (1 sample) and chemical characterization and correlation of ash-fall tephras (21 samples). Broadly these data indicate accumulation of sediment was underway in the lower part of the Skyline Member by 10.27 ± 0.07 Ma (Janecke and Evans, 1999), ceased by ~10.0 Ma in the Cache Valley Member, continued past 5.1–4.4 Ma in the upper part of the Third Canyon Member, and ended perhaps by ~2.0 Ma in the new Canyon Member.

The oldest dated sample is from the 10.27 ± 0.07 Ma Arbon Valley Ash-flow Tuff near the base of the Cache Valley Member. This tuff is a member of the Starlight Formation (Kellogg et al., 1994). Overlying the Arbon Valley Tuff are a number of fine-grained silver gray ash-fall tuffs. These tuffs are all from sources in nearby volcanic fields along the Yellowstone hotspot track (Table 1). Most of these ash-fall tuffs have the general compositional characteritics of ashes produced by ~10.5 to 8.0 Ma explosive eruptions in the Twin Falls/Picabo volcanic fields 200-150 km WNW to NW of the study area (Perkins and Nash, 2002). The youngest hotspot ash-fall tuffs, high in the Third Creek member, are 5.1–4.4 Ma tuffs from sources in the Heise volcanic field centered ~150 km north of the study area.

Based on electron-probe microanalysis of glass shards, about 13 different ash-fall tuffs are recognized in the study area (Table 1). With methods discussed by Perkins et al. (1995; 1998), 8 of these tuffs are correlated with varying degrees of confidence to ash beds dated in areas of the northeastern Basin and Range or the western Snake River Plain. The oldest of these 8 correlative tuffs is in the upper part of the Cache Valley Member. This tuff, collected in four areas, is a possible equivalent of the 10.13 ± 0.13 Ma tuff of Wooden Shoe Butte in the Trapper Creek section of Perkins et al. (1995). Above this tuff are 5 tuffs in the Third Creek member that correlate either with tuffs in the ~10.5 to 6.0 Ma Rush Valley section or the 8.5–6.0 Ma Chalk Hills section of Perkins et al. (1998). These include an unnamed ~9.6 Ma ash bed (Rush Valley sample rv88-4), and unnamed ~8.4 Ma ash bed (sample rv89-9), the ~8.3 Ma Inkom ash bed, and unnamed ~8.1 Ma ash bed (Chalk Hills sample clk93-02), and the ~7.9 Ma Rush Valley ash bed. Age estimates for these 5 tuffs are all interpolation age estimates based on methods discussed by Perkins et al. (1998). The youngest correlative tuffs are the 5.1 ± 0.1 Santee ash bed and the 4.45 ± 0.05 Ma Kilgore ash beds of Perkins and Nash (2002). As discussed by Henry and Perkins (2001) both these ash beds are compositionally similar to basal ash-fall of 4.45 Ma tuff Kilgore in the Heise volcanic field. However, we currently believe compositional, stratigraphic, and isotopic age dating indicate the presence of two different Kilgore-like ash beds in the study area, the older Santee variant matching a dated ash bed in the Great Plains and the Kilgore variant matching the younger tuff of Kilgore.

Finally, the absence of several ash beds in the study area provides indirect evidence of a possible unconformity within the Salt Lake Formation as well as a possible upper age limit for the Salt Lake Formation. The two oldest Heise volcanic field ash beds are the 6.62 ± 0.3 Ma Blacktail Creek ash bed and the 6.29 ± 0.05 Ma Walcott ash bed. These ash beds are present as thick layers in many areas of northwestern Utah (Perkins et al. 1998; Perkins and Nash, 2002), so their absence in the study area suggests a possible unconformity in the upper part of the Third Creek Member between the 7.9 Ma Rush Valley ash bed and the 5.1 Ma Santee ash bed. Also absent in the study area is the 2.06 Ma Huckleberry Ridge ash bed, the very thick and widespread ash bed associated with the eruption of the Huckleberry Ridge Tuff in the Yellowstone Plateau. This ash bed is present in flat-lying strata of the nearby Thatcher basin (Izett and Wilcox, 1982), so its absence in the tilted Salt Lake Formation suggests that the accumulation and deformation of the New Canyon Member predate the eruption of the Huckleberry Ridge Tuff. The abrupt appearance of recycled clasts of the Cache Valley Member in the overlying Third Creek Member (Fig. 9b) suggests that the contact with the underlying Cache Valley member is probably an angular or progressive unconformity. Exposures are too poor to assess the nature of the contact in the field.

The New Canyon Member of the Salt Lake Formation is devoid of ash, but must postdate the 4.4 to 5.1 Ma uppermost portions of the Third Creek Member. If one assumes that the average accumulation rates of 300 m/My determined by Perkins et al. (1998) for Miocene depositional basins in the northern Basin and Range province applies to this part of the Salt Lake Formation, then the top of the roughly 1 km of New Canyon Member could be less than 2 Ma. These accumulation rates match the 290 to 350 m/m.y. rates...
determined in the Lava Hot Springs area for the lower and middle parts of the Salt Lake Formation (Crane, 2000; Kruger et al., this volume). However, the resultant ~2 Ma age estimate for the top of the Salt Lake Formation is very uncertain because accumulation rates for the upper conglomeratic deposits are not known and the original thickness of the New Canyon Member has not been determined. Sediment accumulation rates increase dramatically east of Clarkston Mountain during this time period (Oaks, 2000). If the rates were higher than 300 m/m.y., the top of the New Canyon Member would be older than 2 Ma.

**Interpretation of the Salt Lake Formation**

The depositional basin of the Salt Lake Formation clearly evolved through time (Table 2). The earliest extension is recorded by the Skyline Member in the western part of the Deep Creek half graben (Janecke and Evans, 1999), by the Red Conglomerate in the Cottonwood Valley area (Sacks and Platt, 1985) and by the Collinston Conglomerate in the northern Wellsville Mountains (Goessel, et al., 1999). These presumably correlative conglomerates contain carbonate clasts from underlying Paleozoic rocks and were likely deposited in alluvial fans. In addition, these conglomerates either pinch out within short lateral distances, or, more likely, were localized in small rift basins (episode 1a, Table 2) (Sacks and Platt, 1985; Goessel et al., 1999; Janecke and Evans, 1999).

Tuffaceous beds of the Cache Valley Member lap across the fault-bounded (?) lenses of these basal conglomerates across most of the region and were deposited directly on pre-Tertiary bedrock in the footwalls. Starting before 10.27 ± 0.07 Ma, a broad lake transgressed across the irregular topography formed during the previous minor episode of normal faulting. Internal drainage and arid conditions (Davis and Moutoux, 1998) combined to produce a saline/alkaline lake within the Deep Creek and Long Divide Hills subbasins of the larger Cache Valley Lake (Fig. 10a). Ash from the Yellowstone Hotspot was deposited into freshwater lakes, deltas, beaches and alluvial fans in these half graben. Carbonate deposition occurred within and along the margins of these lakes when the input of ash was reduced. Very similar histories in both the study area and the Oneida Narrows/Cottonwood Valley areas (Danzl, 1985; Sacks and Platt, 1985) probably reflect similar tectonic events in parallel half graben. The Swan Lake tilt block, rocks east of the Clifton Cemetery Fault (and its along-strike continuation), and the proto-Malad Range are examples of fault blocks that formed at this time (Fig. 10b). External drainage developed, and freshwater lakes replaced the saline/alkaline lakes of the prior episode.

Subaerial conditions replaced the lacustrine, and marginal-lacustrine to fluviatile conditions during deposition of the New Canyon Member (Fig. 10c). This thick fluvial unit contains mostly quartzite clasts from Neoproterozoic-Cambrian Brigham Group quartzites and some carbonate and chert clasts from lower Paleozoic formations. The Cache Valley Member had been stripped from the highlands by this time, and provided...
little material to the basins (Figs. 4, 11). Relative rates of subsidence decreased, so that sedimentation continually filled the basins. Deposition and tilting of the Salt Lake Formation probably ended before 2 Ma (see below). The development of the Dempsey Creek half graben, northeast of our study area, followed essentially the same outlines (Crane, 2000; Kruger et al., this volume).

A younger, unrelated (?) period of tectonic activity further disrupted the already dismembered depositional basin of the Salt Lake Formation and allowed the Pliocene-Pleistocene (?) piedmont gravel to prograde across some of the study area.

**Pliocene-Pleistocene (?) Gravel Deposits**

Pliocene-Pleistocene (?) piedmont gravel deposits are the principal sedimentary record of younger Basin-and-Range faulting (episode 3). These are preserved as local, dissected erosional remnants within the Clifton Horst, and as widespread piedmont deposits in two areas west of the horst (units QTg and QTrg in Fig. 3). This sequence is very thin relative to the more than 2 km of Salt Lake Formation. Within the horst, less than tens of meters of the piedmont gravels are preserved, whereas west of the horst more than 200 m are exposed (Fig. 7). Within the Clifton Horst the Pliocene-Pleistocene (?) piedmont gravel form widely scattered erosional remnants around the margins of the highest peaks and as partially exhumed fill of the adjacent mountain valleys. Exposures are concentrated between Davis Basin and the Baldy Peak area (Figs. 3, 8). The lithology, clast size, composition, and degree of cementation of the gravel-and-conglomerate deposits depend strongly on the lithology of the adjacent and underlying bedrock. Around Davis Basin brecciated and re-cemented blocks up to 3 m across of the underlying Brigham Group quartzites armor the hillsides where they are weathering out of this unit. A short distance to the south, in Clifton Basin, a small, well-cemented, erosional remnant of the Pliocene-Pleistocene (?) piedmont gravel deposit lies in angular unconformity on steeply southeast-tilted Salt Lake Formation (Figs. 3, 7e, 8). Here the Pliocene-Pleistocene (?) piedmont gravel contains mostly angular clasts of light colored quartzites derived from the Camelback Mountain Quartzite, and rare, mostly smaller clasts of rust-colored debris from the Pocatello Formation (Fig. 7e). The sources of these lithologies are upslope to the west and north of this erosional remnant. The presence of sediment derived from the Neoproterozoic Pocatello Formation is unique to the Pliocene-Pleistocene (?) piedmont gravel deposits. Such sediment is notably lacking from the Salt Lake Formation (Figs. 4, 11).

The Pliocene-Pleistocene (?) piedmont gravel deposits lie at elevations of ~2000 m to ~2200 m within the Clifton Horst and define irregular bodies of sediment that slope gently eastward or westward from the crest of the range. Correlative deposits are lacking directly east of the horst, and occur at least 100 m lower in elevation west of the horst. These data suggest that there has been more slip across the Dayton-Oxford Fault since deposition of these deposits than across the Deep Creek Fault. The steep rugged topography of the east side of the Clifton Horst supports this interpretation (Fig. 7a).

West of the Clifton Horst an erosionaly dissected, mildly faulted and folded piedmont surface occupies the hanging wall of the Deep Creek Fault (Figs. 3, 7b, 7c, 7d). Small east- and west-dipping normal faults offset and fold the piedmont surface (Fig. 3). These faults strike northerly and are interpreted as synthetic and antithetic faults to the Deep Creek Fault (Figs. 5, 6).

The composition of the gravel-and-conglomerate deposits beneath the piedmont surface varies from north to south and from east to west. West of the Clifton Horst, along the mountain front, the composition of the clasts in the pebble-to-cobble gravels and conglomerates reflects the lithology of the bedrock immediately to the east in the Clifton Horst. For example, the Pliocene-Pleistocene (?) piedmont gravel directly west of Weston Peak contain angular pebbles of quartzite from the Brigham Group and is not cemented (Fig. 3). Just 1.2 km to the south along stratigraphic and structural strike, the conglomerates consist of gently west-sloping beds of limestone-clast conglomerate that contain identifiable pebbles of the Blacksmith Formation, which is exposed 0.6 km to the east in the footwall of the Deep Creek Fault. The gravels and conglomerates along the mountain front are interpreted as alluvial-fan deposits shed from the western edge of the Clifton Horst.

In a second area of exposures, the younger Pliocene-Pleistocene (?) piedmont gravel deposits overlie folded beds of the Cache Valley Member southeast of Dry Creek (Fig. 3). There The New Canyon and Third Creek Members were removed by erosion prior to deposition of these gravels. Well-rounded cobbles and boulders of quartzite that dominate the deposit were recycled from the New Canyon Member of the Salt Lake Formation, which is exposed upslope to the east and northeast. Several north-striking normal faults cut...
this Pliocene-Pleistocene (?) piedmont gravel deposit, and produced a broad south-plunging rollover anticline in their hanging walls. Exposures are poor, but at least 100 m of Pliocene-Pleistocene (?) piedmont gravels and conglomerates are preserved here. Compositions of axial deposits suggest north to south transport of these gravels in the Deep Creek Half graben.

The Pliocene-Pleistocene (?) piedmont gravel deposits overlie Paleozoic and Neoproterozoic bedrock in most exposures within the Clifton Horst, and overlie the Salt Lake Formation west of the horst. The contact between the Pliocene-Pleistocene (?) piedmont gravel deposits and the pre-Tertiary units is everywhere an angular unconformity. The nature of the contact between the Pliocene-Pleistocene (?) piedmont gravel deposits and the older Salt Lake Formation is less clear in the field, but map relationships indicate an angular unconformity or disconformity there.

Similar Relationships In Neighboring Areas

Dissected piedmont surfaces, erosion surfaces, active pediment surfaces, and associated gravel deposits of variable thickness are common in southeastern Idaho and northern Utah (Ore, 1982; Oaks et al., 1999; Ore, 1999; Kruger et al., this volume). Erosional truncation of the western margin of the Deep Creek half graben, for example, planed off all but the highest peaks above 2100 m. Northward, the Tertiary basin fill in the Malad Summit basin is truncated by inward-sloping piedmont surfaces that are graded to a level above the current valley floor. High surfaces in northwestern Malad Valley and along the margins of Marsh Valley have a similar origin (Fig. 2). Piedmont gravel deposits and erosion surfaces are less common along the margins of Cache Valley because Cache Valley was more active in the last few million years and because it was extensively modified by Lake Bonneville. In Cache Valley, piedmont gravel deposits and erosion surfaces are restricted to up-faulted blocks of basin fill along the margins of the valley, southeasternmost Cache Valley, in the northern Wellsville Mountains and on the southern flank of Clarkston Mountain (Oaks et al., 1999; Goessel, 1999; Oaks, 2000). In Utah the piedmont deposits slope valleyward, are slightly faulted and folded, like those in the Deep Creek half graben, and are interpreted as a record of tectonic quiescence and stable base level prior to the integration of the Cache Valley drainage system into the larger drainage basin of the Great Salt Lake (Oaks et al., 1999; Oaks, 2000).

Overall, there was little sedimentation in greater Cache Valley area after the end of deposition of the Salt Lake Formation. This is indicated by the thin Pliocene-Pleistocene (?) piedmont gravel deposits and by subsurface data. Well logs from drill holes show that there are only a few hundred meters of sediment preserved beneath Cache Valley above the tilted tuffaceous units of the Salt Lake Formation, except in the area along the central segment of the East Cache Fault (Williams, 1962; Robinson, 1999). The only other possibly thick accumulation of sediment from this time period is a potentially very young sequence of sand, silt and gravel 915 to 1220 m thick within the Washboards subunit of the Salt Lake Formation northeast of Clarkston Mountain (Oaks, 2000; Biek et al., 2001). However, Oaks (2000) and Biek et al. (2001) tentatively interpreted these uplifted exposures as part of the Salt Lake Formation, in part because deposits are tilted, conformable with the Salt Lake Formation and cut by a discordant erosion surface (Biek et al., 2001; Oaks, 2000). Perhaps, less accommodation space was created in the Cache Valley area during Basin-and-Range faulting than during the prior episode of detachment faulting.

Age of the Pliocene-Pleistocene (?) Piedmont Gravel Deposits

The age of the Pliocene-Pleistocene (?) piedmont gravel deposits in the area of study is poorly constrained between 5.1 or 4.4 Ma (the youngest correlated age from the Salt Lake Formation) and late Pliocene deposits of Lake Bonneville (about 25 to 15 ka). Extrapolating typical accumulation rates to the top of the Salt Lake Formation suggests that the Pliocene-Pleistocene (?) piedmont gravel deposits might be entirely Quaternary. This inference is consistent with the presence of 0.76 to 0.77 Ma and 0.61 Ma ashes in old alluvial fans that accumulated on a bajada in the northeast corner of modern Cache Valley in angular unconformity on tilted Paleozoic bedrock and Salt Lake Formation (Bright and Ore, 1987; age of Bishop Ash was revised according to Sarna-Wojcicki and Pringle, 2000), and the presence of flat-lying 2 Ma Huckleberry Ridge tuff in southern Gem Valley (located at the star in Fig. 2) (Izett, 1982; Bouchard et al., 1998). In southern Cache Valley, McCalpin (1994) estimated an early to mid Pleistocene age of the elevated and faulted pediment gravels there.

In Marsh Valley a flat-lying 0.583 ± 0.104 Ma basalt flow (Scott, 1982) appears to fill a valley inset into piedmont deposits that resemble the Pliocene-Pleistocene (?) gravel deposits. This suggests that the piedmont gravels in Marsh Valley are late Pliocene to mid Pleistocene in age. If the flat attitude of the
Huckleberry Ridge Tuff in Gem Valley is representative of the entire region, it may indicate that tilting in the Salt Lake Formation to angles up to 90° (Sacks and Platt, 1985; Janecke and Evans, 1999) was completed before 2 Ma. Tilting was certainly completed before the older alluvial fans that contain the 0.76 to 0.77 Ma Bishop Ash (Bright and Ore, 1987) were deposited in northern Cache Valley.

Interpretation of the Pliocene-Pleistocene(?) Piedmont Gravel

The present high elevation of the Pliocene-Pleistocene(?) piedmont gravel deposits within the Clifton Horst, and their relationships with underlying Tertiary to Neoproterozoic deposits, suggest that uplift of the horst occurred during at least two pulses of activity separated by a period of relative tectonic quiescence (Table 2). Facies patterns show that the Salt Lake Formation was once continuous across the present position of the Clifton Horst. Erosion of the 2 to 3 km thickness of Salt Lake Formation, the underlying Paleozoic rocks (up to 2.5 km), and the Neoproterozoic rocks (up to 3 km) began when the Clifton Horst was first uplifted along the north-striking Dayton-Oxford and Deep Creek normal faults some time after 4.4 or 5.1 Ma (episode 3a, Table 2; Fig. 6). Erosion had stripped the horst to near its present erosion level when a period of tectonic quiescence allowed alluvial fans and hillslope deposits to nearly bury the horst in its own debris (episode 3b, Table 2). Only the highest peaks remained as inselbergs above the piedmont surface. A long period of tectonic stability is indicated by the presence of this sediment in the interior portions of the Clifton Horst in canyons between some of the highest peaks in the range (Fig. 3).

Renewed slip on the Dayton-Oxford normal fault elevated these alluvial-colluvial deposits at least 425 m above correlative strata in northernmost Cache Valley and separated them from their downslope equivalents (episode 3c, Table 2). Other lateral equivalents may lie buried beneath Cache Valley, or were removed by erosion. Slip on the Deep Creek Fault, on the western edge of the horst, was significantly less than that along the eastern margin of the horst during this youngest episode of normal faulting. The piedmont surfaces there were offset approximately 100 m during this episode of slip. Small normal faults antithetic to the Deep Creek Fault were activated, and the Pliocene-Pleistocene(?) piedmont gravel deposits were locally folded. A period of active uplift of the horst, followed by a tectonic quiet period and then renewed tectonism likely explains the piedmont gravel deposits.

A similar quiescence in tectonism has been interpreted to have affected the Utah portion of Cache Valley at essentially this same time, after deposition of the Salt Lake Formation (after 4.1 or 5.1 Ma) but before integration of the drainage system in Cache Valley with the larger drainage basin of the Great Salt Lake Valley (Oaks et al., 1999; Oaks, 2000). Pediment surfaces in the Utah portion of Cache Valley formed during this tectonic lull (Oaks et al., 1999). Improved geochronology is needed to determine whether the lull was synchronous across the region.

The significance of this tectonic lull and the increased activity that followed is obscure. A period of suppressed tectonism is ongoing evident today in the region immediately to the north in Marsh Valley and its environs, within neotectonic Belt IV of Pierce and Morgan (1992) (Fig. 1). That region has well-developed, high-level piedmont surfaces which probably predate the poorly dated 0.583 ± 0.104 Ma Portneuf basalt flow (Scott, 1982), and has experienced little surface faulting since late Pliocene time (Pierce and Morgan, 1992). Gravity data show a deep basin in the southern end of Marsh Valley that might have formed during a vigorous interval of Basin-and-Range faulting before the high level piedmont surfaces formed. If the Pliocene-Pleistocene(?) period of tectonic stability in Cache Valley was related to the ongoing quiescence in neotectonic Belt IV, understanding its origin will place additional constraints on models of Basin-and-Range extension in the wake of the northeast-migrating Yellowstone Hotspot.

Evolution of the Rift Basins

Together the structural and sedimentologic data suggest the following evolution. Small rift basins filled with alluvial-fan deposits were scattered across the region between 12 and 10.5 Ma (Episode 1a, Table 2). Initiation of the Bannock detachment system produced the original, broad depositional basin occupied by the Cache Valley Member of the Salt Lake Formation. Simple translation characterized the hanging wall of the Bannock detachment system at this time (Episode 1b, Table 2; Fig. 10a). The broad basin was disrupted at least twice by normal faults (Table 2). The first structural “break-up” (episode 1c; Fig. 10) occurred in the hanging wall of west-southwest-rooted, low-angle normal faults of the Bannock detachment system, whereas a second younger disruption (episode 3) coincided with the development of unrelated steeper range-front faults along the
An incompletely understood episode of faulting occurred along southeast, east- and northeast-striking normal faults (episodes 3a to 3c). Minor sedimentation formed the Pliocene-Pleistocene piedmont gravel deposits during the younger development of the modern topography. Minor sedimentation formed the Pliocene-Pleistocene (?) piedmont gravel deposits during the younger development of the modern topography along north-striking normal faults (episodes 3a to 3c). A drop in base level accompanied the development of the modern topography of Cache Valley sometime during episode 3 (e.g., Oaks et al., 1999) and might explain the sparse sedimentary record of this time interval.

DISCUSSION

The northeast corner of the Great Basin, like many other parts of the Basin-and-Range province, formed during at least two episodes of roughly east-west extension. An early episode of extension produced low-angle normal faults and basins and highland-like the modern ones. A later episode of Basin-and-Range extension overprinted the earlier deformation and generated the modern topography. Multiple episodes of extension in the Basin-and-Range province are well characterized (Proffett, 1977; Taylor et al., 1989; Stewart, 1998; Janecke et al., 2001) but the pattern in southeast Idaho and north-central Utah is unusual because large magnitude extension began so recently (<12 Ma) and may have continued into the late Pliocene.

The Bannock detachment system projects northward to the Snake River Plain (Fig. 1) (Janecke and Evans, 1999; Kruger et al., this volume). The inferred listric breakaway, the Valley Fault, is the most continuous structure in the system and probably persists at least to the latitude of Pocatello based on interpretation of relationships in Link and Stanford (1999). Tilted and faulted Salt Lake/Starlight Formation in the hanging wall of the Valley Fault (and along strike equivalents) persist along the axis of the Portneuf Range into the northern part of the range (Link and Stanford, 1999; Crane, 2000; Crane et al., 2001; Kruger et al., this volume) in accordance with this interpretation (Fig. 1). The continuity of the low-angle part of the fault system in the subsurface and northward is less certain because younger normal faults offset it (Link and Stanford, 1999; c.f., Kruger et al., this volume), and because the pre-extensional geometry of the region is incompletely known. The Bannock detachment system may connect with low-angle normal faults that overlie the Pocatello Formation north of Hawkins Creek (north edge of Fig. 2) (Platt, 1985) and extend northward toward Pocatello. There the Fort Hall normal fault and its continuation may wrap around the northern end of the Portneuf Range (based on map data in Trimble, 1976; Rodgers and Othberg, 1996; Kellogg et al., 1999; Rodgers et al., in press) and have a geometry like the folded Clifton Fault of the Bannock detachment system in the Clifton Horst. Overturned Pocatello Formation in the Pocatello Range is localized in the footwall of the low-angle fault system according to this interpretation. Further work is needed to test this interpretation hypothesis and the alternative in Rodgers et al. (in press).

The Bannock detachment system probably continues in modified form to the southernmost parts of Cache Valley if the continuity of the Salt Lake Formation and its similar ages and facies (Oaks et al., 1999) are used as an indicator (Fig. 2). Between Oneida Narrows and a few kilometers north of Logan, the eastern strand of the listric East Cache Fault dips as low as 22° at the surface (Fig. 2) (Brummer, 1991; Dover, 1995). Farther south, seismic-reflection data show that the eastern strand of the East Cache Fault zone steepens progressively southward, but it retains a listric geometry near Hyrum and southward at least to the south end of Cache Valley (Evans and Oaks, 1996) (Fig. 2). Thus, the fairly inactive eastern strand of the East Cache Fault zone from Logan northward likely represents the along-strike continuation of the breakaway of the Bannock detachment system, whereas the western strands of the East Cache Fault zone in the same area sustained most of the younger Basin-and-Range extension. Gravity data are consistent with this
been documented in several parts of the Basin and dipping normal faults in less extended areas have operated during deposition of the Cache Valley Member.

The amount of extension during episode 1 probably increases northward from the south end of Cache Valley, where the extensional system consists of a single half-graben (southernmost Cache Valley) and the relatively unfaulted Wellsville tilt block to the west (Smith, 1997; Oaks et al., 1999). Numerous north-striking normal faults first appear in the northernmost Wellsville Mountains (Goessel et al., 1999) and typify Clarkston Mountain (Biek, 1999a; Biek et al., 2001), the Malad Range (Janecke and Evans, 1999), Oxford Ridge (Link, 1982a and b; this study) (Fig. 2), and Elkhorn Mountain (Oriel et al., 1991; Pope and Link, 2000; Pope et al., 2001). Low-angle normal faults, which can accommodate greater extension than high-angle normal faults, are present from Junction Hills northward to Marshville tilt block to the west (Smith, 1997; Oaks et al., 1999). Numerous north-striking normal faults first appear in the northernmost Wellsville Mountains (Goessel et al., 1999) and typify Clarkston Mountain (Biek, 1999a; Biek et al., 2001), the Malad Range (Janecke and Evans, 1999), Oxford Ridge (Link, 1982a and b; this study) (Fig. 2), and Elkhorn Mountain (Oriel et al., 1991; Pope and Link, 2000; Pope et al., 2001). Low-angle normal faults, which can accommodate greater extension than high-angle normal faults, are present from Junction Hills northward to Marshville tilt block to the west (Smith, 1997; Oaks et al., 1999). Numerous north-striking normal faults first appear in the northernmost Wellsville Mountains (Goessel et al., 1999) and typify Clarkston Mountain (Biek, 1999a; Biek et al., 2001), the Malad Range (Janecke and Evans, 1999), Oxford Ridge (Link, 1982a and b; this study) (Fig. 2), and Elkhorn Mountain (Oriel et al., 1991; Pope and Link, 2000; Pope et al., 2001). Low-angle normal faults, which can accommodate greater extension than high-angle normal faults, are present from Junction Hills northward to Marshville tilt block to the west (Smith, 1997; Oaks et al., 1999). Numerous north-striking normal faults first appear in the northernmost Wellsville Mountains (Goessel et al., 1999) and typify Clarkston Mountain (Biek, 1999a; Biek et al., 2001), the Malad Range (Janecke and Evans, 1999), Oxford Ridge (Link, 1982a and b; this study) (Fig. 2), and Elkhorn Mountain (Oriel et al., 1991; Pope and Link, 2000; Pope et al., 2001). Low-angle normal faults, which can accommodate greater extension than high-angle normal faults, are present from Junction Hills northward to Marshville tilt block to the west (Smith, 1997; Oaks et al., 1999). Numerous north-striking normal faults first appear in the northernmost Wellsville Mountains (Goessel et al., 1999) and typify Clarkston Mountain (Biek, 1999a; Biek et al., 2001), the Malad Range (Janecke and Evans, 1999), Oxford Ridge (Link, 1982a and b; this study) (Fig. 2), and Elkhorn Mountain (Oriel et al., 1991; Pope and Link, 2000; Pope et al., 2001). Low-angle normal faults, which can accommodate greater extension than high-angle normal faults, are present from Junction Hills northward to Marshville tilt block to the west (Smith, 1997; Oaks et al., 1999). Numerous north-striking normal faults first appear in the northernmost Wellsville Mountains (Goessel et al., 1999) and typify Clarkston Mountain (Biek, 1999a; Biek et al., 2001), the Malad Range (Janecke and Evans, 1999), Oxford Ridge (Link, 1982a and b; this study) (Fig. 2), and Elkhorn Mountain (Oriel et al., 1991; Pope and Link, 2000; Pope et al., 2001).
Figure 10 (opposite page and this page): Reconstructed paleogeographic maps and cross sections of the northern Cache Valley area between the modern West Hills/Samaria Mountain and the Bear River Range. Extension was reconstructed based on a preliminary reconstruction of Figure 5. Episodes of extension correspond to those in Table 2. Episodes 1a, 2, and 3a are not illustrated because their paleogeography is incompletely known.

A) Initiation of slip on the detachment system after about 10% extension (50% extension was restored). Internal drainage prompted the development of saline/alkaline lake conditions in at least two sub-basins of Cache Valley Lake. Faults that developed during the next episode are shaded. B) Early during Episode 1c. The hanging wall of the Bannock detachment system broke up internally along southwest-dipping normal faults, smaller half graben formed, uplifted Cache Valley Member was recycled into younger deposits, and external drainage developed. Freshwater lakes and nearshore deposits formed. C) Late during Episode 1c. Exhumation of tilted blocks exposed more pre-Tertiary bedrock. Braided streams filled the Deep Creek sub-basin. External drainage. The future positions of the Clifton Horst and the Elkhorn Mountain-Clarkston Mountain fault blocks are outlined with dashed lines. D) Slip on southwest-dipping Bannock detachment system ceased. High-angle normal faults initiated. These dismembered the rift basins of the Salt Lake Formation. Piedmont gravel buried the Basin-and-Range topography during a tectonic lull. By the end of episode 3, external drainage was well established. See text for more detail.
CONCLUSIONS

Geologic studies in the 7.5-minute Clifton and Malad City East quadrangles provides evidence for multiple episodes of westward extension in southeast Idaho during deposition of the late middle Miocene to Pliocene Salt Lake Formation and younger Pliocene-Pleistocene(?), piedmont gravels. North-northwest trending, south-southwest dipping detachment faults and younger associated closely spaced normal faults in their hanging wall accommodated the earliest large-magnitude extension. The close spatial association between the highly extended terrane of the Bannock detachment system and the Cache-Pocatello Culmination of the Cordilleran fold-and-thrust belt, suggest that gravitational collapse likely contributed to this main episode of extension. Other thermo-tectonic factors might also have triggered the extensional collapse. Tectonic models that relate the extension during episode 1 to passage of the Yellowstone Hotspot alone must be in error.

South of the lateral margin of the Cache-Pocatello Culmination, the Bannock detachment system steepens, displacement decreases, and deposition of the Salt Lake Formation was confined to a single, deep half-graben in the southern end of Cache Valley (Oaks, et al., 1999). The Miocene-Pliocene Salt Lake Formation that was deposited in this area of study shows evidence for changing depositional systems from fluvial to under-filled lacustrine to overfilled fluvial in response to the coalescing and then fragmenting extensional basins. Saline/alkaline lakes formed early during episode 1 and altered much of the voluminous, reworked ash to clinoptilolite. The sedimentary basins of the Salt Lake Formation were larger than their modern remnants, and have been reconstructed across younger fault blocks of the Cache Valley region.

Pliocene-Pleistocene(?), piedmont gravels are the thin sedimentary record of the modern Basin-and-Range extensional system. Little sediment is preserved in and around Cache Valley from the time of Basin-and-Range extension (episode 3), possibly due to integration of the drainage basin into the rift basins of the Great Salt Lake. A protracted(?) period of tectonic quiescence during episode 3 allowed piedmont gravel deposits to bury all but the highest peaks of Clifton Horst, and permitted pediments to form elsewhere around the region. Subsequent renewed uplift elevated these piedmont deposits in the Clifton Horst at least 425 m above their equivalents to the east in Cache Valley. Marsh Valley, due north of Cache Valley, may preserve a more substantial sedimentary record of Basin-and-Range extension (Kruger et al., this volume) but is currently less active than Cache Valley (Pierce and Morgan, 1992).

This study illustrates a common sequence of structural-stratigraphic events in the Basin-and-Range province: 1) Early, large-magnitude extension along a listric low-angle normal fault; 2) break-up of the hanging wall of the low-angle normal fault along numerous normal faults; and 3) subsequent small-magnitude extension along steeper, widely spaced range-front faults that dismember the earlier detachment system (see also Proffett, 1977). However, the entire sequence of events is anomalously young in the Cache Valley region with initial extension beginning in latest Middle Miocene to early Late Miocene time, and emergence of the modern basins and ranges in mid to late Pliocene. An incompletely characterized episode of cross faulting, possibly due to flexure toward the Eastern Snake River Plain, further complicated the extensional history of the greater Cache Valley region during episode 2. This evolutionary sequence may be a natural cycle of extensional deformation in gravitationally collapsing regions.

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