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Evolution of a Miocene-Pliocene Low-Angle Normal-Fault System in the Southern Bannock Range, Southeast Idaho

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EVOLUTION OF A MIOCENE-PLIOCENE LOW-ANGLE NORMAL-FAULT SYSTEM IN THE SOUTHERN BANNOCK RANGE, SOUTHEAST IDAHO

by

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ABSTRACT

Evolution of a Miocene-Pliocene Low-Angle Normal-Fault System in the Southern Bannock Range, Southeast Idaho

by

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Utah State University, 2002

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Geologic mapping, basin analysis, and tephrochronologic analysis in the Clifton quadrangle of southeast Idaho indicates that the modern Basin-and-Range topography is only a few million years old and that the bulk of Cenozoic extension was accommodated by slip on an older low-angle normal-fault system, the Bannock detachment system. The detachment system was active between ~12 and < 4 Ma and accommodated ~50 % extension.

Cross-cutting relationships show that the master detachment fault, the Clifton fault, is the youngest low-angle normal fault of the system, was active at a low angle, and has not been rotated to a low-dip angle through time. Map patterns and relationships indicate that the hanging wall to the detachment system began as a cohesive block that later broke up along listric and planar normal faults that either sole into or are cut by the master detachment fault. The Miocene-Pliocene Salt Lake Formation, a syntectonic, basin-fill deposit of the Bannock detachment system, was deposited during three sub-episodes of extension on the detachment system. Depositional systems within the Salt Lake Formation evolved from saline/alkaline lakes to fresh water lakes and streams to braided streams in response to the changing structural configuration of rift basins in the hanging wall of the detachment system. After breakup of the hanging wall began, the master detachment fault excised part of the hanging wall and cut hanging-wall deposits and structures.

The structural geometry of the Bannock detachment system strongly resembles that of detachments documented in metamorphic core complexes. Therefore, we interpret the Bannock detachment system as a proto-metamorphic core complex, akin to the Sevier Desert detachment fault. The Bannock detachment system also collapsed the Cache-Pocatello culmination of the dormant Sevier fold-and-thrust belt, much like the Sevier Desert detachment collapsed the Sevier culmination.

Structures of the Bannock detachment system are overprinted by a second episode of extension accommodated by E- and NE-trending normal faults that may be related to subsidence along the Yellowstone hotspot track and a third episode of extension accommodated by high-angle, Basin-and-Range normal faults. This last episode of extension began no earlier than 4-5 Ma and continues today.

(177 pages)

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Stephanie M. Carney

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CHAPTER 1

INTRODUCTION

A system of NNW-striking, low-angle normal faults is exposed in the southern Bannock Range of southeast Idaho. Early mapping and structural studies in this area documented younger-on-older normal faults (interpreted as younger-on-older thrusts by Raymond [1971]) that, in some areas, juxtapose Cambrian strata against Neoproterozoic rocks (Murdock, 1961; Raymond, 1971; Mayer, 1979; Link, 1982a, 1982b; and Oriel, unpublished). These studies also documented that these low-angle faults were older than the modern Basin-and-Range topography. More recent work by Janecke and Evans (1999) also documented these low-angle normal faults in the southern Bannock Range and showed that the Tertiary Salt Lake Formation is the youngest unit in low-angle fault contact with the Neoproterozic Pocatello Formation. They suggest that the low-angle normal faults were active in the Tertiary and that the Miocene-Pliocene Salt Lake Formation is syntectonic, deposited during slip on the low-angle normal faults. They named the system of low-angle normal faults the Bannock detachment system and suggested that the Valley fault of Sacks and Platt (1985), farther east in the Portneuf Range, is a possible breakaway for the detachment system.

The Salt Lake Formation and its equivalent, the Starlight Formation, are widely exposed in northern Utah and southern Idaho. These rocks have been studied in detail in this area by many workers (Adamson et al., 1955; Danzl, 1985; Sacks and Platt, 1985; Smith, 1997; Goessel, 1999; Goessel et al., 1999; Janecke and Evans, 1999; Oaks et al., 1999; Crane, 2000; Oaks, 2000) and have long been interpreted as the sedimentary record of Basin-and-Range extension that is thought to have begun about 17 Ma in the Basin-and-Range province (Williams, 1948; Slentz, 1955; Hintze, 1988; Bryant et al., 1989; Stewart, 1998). Preliminary geochronologic studies showed that the Salt Lake Formation is middle to late Miocene-early Pliocene in age (Smith and Nash, 1976; Smith, 1997; Goessel, 1999). More recent studies (Janecke and Evans, 1999) indicate that deposition of the Salt Lake Formation is linked to the older Bannock detachment system, and not younger Basin-and-Range extension. The older Bannock extensional system is characterized by more extension and the presence of low-angle normal faults.

Low-angle normal-fault (detachment-fault) systems are regionally extensive, accommodate large amounts of horizontal strain, and are, therefore, important structures in extending terranes (Allmendinger et al., 1983; Wernicke, 1995). Low-angle normal faults are widely recognized in the Basin-and-Range province of the western United States and are most commonly associated with highly extended metamorphic core complexes in the North American Cordillera (Longwell, 1945; Armstrong, 1972; Proffett, 1977; Allmendinger et al., 1983; Davis and Lister, 1988). Much work has been done in documenting the structural and stratigraphic relationships of normal faults and syntectonic basin-fill deposits associated with these detachment faults. However, there is much controversy over whether or not normal faults can accommodate slip at dip angles less than 30° with reasonable pore pressures (Wernicke et al., 1985; Spencer, 1985; John, 1987; Yin, 1989; Brady et al., 2000). Few earthquakes have been identified on low-angle normal faults (Jackson and White, 1989; Abers 1991; Boncio et al., 2000) and classic Andersonian mechanics does not predict slip on normal faults with dips less than 30° (Anderson, 1951).

Detachment faults and listric normal faults have also been documented as the main structures that collapse thrust-related culminations in the Cretaceous-early Tertiary Sevier fold-and-thrust belt of the western United States (Bryant, 1990; Constenius, 1996; Yonkee, 1997; Janecke et al., 2001). Work by Rodgers and Janecke (1992) indicated that a large, broad, uplifted area is present in southeastern Idaho in the hanging wall of the Paris-Putnam thrust. This uplifted area is better defined by this study and is termed a culmination. This culmination seems to coincide in area with the Bannock detachment system in southern Idaho and north-central Utah.

The purposes of this study are: 1) to better understand the structural and stratigraphic evolution of the Salt Lake Formation as a syntectonic deposit coeval with slip on the Bannock detachment system, and 2) to perform a detailed structural analysis of the Bannock detachment system in order to determine its structural evolution and its bearing on the low-angle-fault controversy, how it compares to other low-angle normal-fault systems, and its relationship to the thrust-related, uplifted area described by Rodgers and Janecke (1992).

Building on unpublished work in the 1970s by the late Steven S. Oriel, geologic mapping of the Clifton 7.5-minute quadrangle (Plate 1) in the southern Bannock Range of southeast Idaho provides key evidence that sheds light on how the Bannock detachment system and syntectonic deposition of the Salt Lake Formation evolved during the Miocene-Pliocene. Chapter 2 focuses on the deposition of the Tertiary Salt Lake Formation. The lithology, stratigraphy, and depositional environments of the Salt Lake Formation are examined in detail in order to determine the paleogeography of the northern Cache Valley area and to shed light on the interplay of sedimentation and tectonics within evolving rift basins. In this study I build on the initial stratigraphic and sedimentologic work on the Tertiary Salt Lake Formation by Janecke and Evans (1999) and propose a regional interpretation of a depositional history controlled by extension on the Bannock detachment system. Chapter 3 focuses on the structural evolution of the Bannock detachment system. The structural characteristics and geometric relationships of the low-angle faults of the Bannock detachment system, as well as hanging-wall and footwall structures, are closely examined and analyzed in order to determine the origin and evolution of this detachment system and its bearing on of the controversy about the original dip of low-angle normal faults. Structural analysis shows that two large, NNW- and ESE-trending anticlines are associated with extension on this detachment system. We examine the possibility that the second of these two folds is the proto-core of the Bannock detachment system and show that the Bannock detachment system may be a proto-metamorphic core complex of Miocene-Pliocene age. The construction of cross sections of the geologically mapped Clifton quadrangle, as well as a cross section constructed from a compilation of regional mapping, allowed the amount of extension accommodated by the Bannock detachment system and the Basin-and-Range faults to be determined. A subcrop map that expands on the work of Rodgers and Janecke (1992) was constructed to help determine the relationship of the Bannock detachment system to older structures of the Sevier Orogeny.

Chapter 2 is modified slightly from Janecke et al. (in press), which was submitted in May 2002 for publication in 'Cenozoic Paleogeography of the Rocky Mountains', a special publication of the Society for Sedimentary Geology. Chapter 3 will be submitted to the Geological Society of America Bulletin. The geologic map of the Clifton quadrangle (Plate 1) will be submitted to the Idaho Geological Survey for publication in their Technical Reports series.

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CHAPTER 2

LATE MIOCENE-PLIOCENE DETACHMENT FAULTING AND PLIOCENE-RECENT BASIN-AND-RANGE EXTENSION INFERRED FROM DISMEMBERED RIFT BASINS OF THE SALT LAKE FORMATION, SOUTHEAST IDAHO[1]

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 formed during offset on 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 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

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[1] Coauthored by Stephanie M. Carney, Susanne U. Janecke, Michael E. Perkins, Jeffrey C. Evans, Paul K. Link, Robert Q. Oaks, Jr., and Barbara P. Nash.

(<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 and ending after 5.1 or 4.4 Ma) and of the modern Basin-and-Range topography (developing after 4.4 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 (Link and Stanford, 1999), 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. 2-1) (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 (Dumitru 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) suggests 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 and 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 northeastward, 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. 2-2). 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 the Salt Lake Formation along the northeast and east margin of the area in Figure 2-2. 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 (cf., 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 this area will be presented in a separate paper (Carney and Janecke, Chapter 3).

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. 2-2, 2-3) and proposes a regional interpretation based on our new data and previous results from around the northern Cache Valley. 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.

METHODS

The structural evolution and paleogeography of the southern Bannock Range is based on: 1) detailed geologic 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. 2-3); 2) photogeologic mapping on 1:16,000-scale aerial photographs; 3) stratigraphic and sedimentologic analysis of Cenozoic basin-fill deposits (Fig. 2-4, Appendix B); 4) age determinations based on chemical correlations of tephras within the Salt Lake Formation with tephras of known ages (Table 2-1); 5) geologic cross sections (Figs. 2-5, 2-6); and 6) argon geochronology. Mapping in the Clifton quadrangle builds on and revises unpublished mapping by Steven S. Oriel. The Clifton and Malad City East quadrangles contain the Malad Range, Clifton Horst, Deep Creek half graben, and northwest edge of Cache Valley (Figs. 2-2, 2-3). A regional cross section 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. 2-5). 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 extremely 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 2-4 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 2-1). Chemical compositions of 20 glass shards were determined for each sample on the electron microprobe at the University of Utah (Table 2-1) (methods of Perkins et al., 1995, 1998).

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 near the northeastern margin of the Basin-and-Range province (Figs. 2-1, 2-2). The study area is in the hanging wall of the west-dipping Paris-Willard thrust fault, which is exposed approximately 45 km east of the study area, and was active from the Early to Late Cretaceous (Mansfield, 1927; Royse et al., 1977; Wiltschko 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. 2-1) (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 who incorrectly showed Salt Lake Formation in depositional contact with the Neoproterozioc Pocatello Formation) 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 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.

The second episode of extension is characterized by moderately to steeply dipping, planar to listric normal faults that typify the 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. 2-2, 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. 2-2). The modern 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. 2-5, 2-6, 2-7). The northeast margin of Cache Valley is defined by several older, presently inactive, northwest-striking, southwest-dipping normal faults (Fig. 2-2).

OVERVIEW OF STRUCTURAL GEOLOGY

A full description of the structural geology of the area is beyond the scope of this paper but is the subject of a separate effort (Carney and Janecke, Chapter 3). A 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 by several younger episodes of normal faulting (Table 2-2). We separate normal faults in this area into sets of normal faults, or 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 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. 2-8) 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. 2-6). Structural arguments beyond the scope of this paper (Janecke and Evans, 1999; Carney and Janecke, Chapter 3) 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. However, exposures of these faults are limited to the Clifton Horst and Little Mountain area (Figs. 2-2, 2-5). We interpret faults of subset a as normal faults that formed in the hanging wall of detachment faults of subset b (Fig. 2-6). 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. 2-5, 2-6).

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 2-2; Fig. 2-3). 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- and 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. 2-2, 2-3) and 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. 2-7b and 2-7d).

The following observations support our grouping of normal faults into three geometrically and temporally distinct sets; 1) Most normal faults of 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. 2-6); 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 normal faults of sets 1 and 3 have similar strikes (Figs. 2-2, 2-3) (Goessel et al., 1999; Oaks et al., 1999; Kruger et al., in press); 4) Map and cross sectional analysis suggests that some of the west-southwest-dipping normal faults of fault set 1 (subset a) merge with laterally continuous low-angle normal faults of the Bannock detachment system (fault set 1, subset b) in a geometry that typifies highly extended terranes above detachment faults (Figs. 2-3, 2-5, 2-6) (Wernicke, 1992). This association between closely spaced hanging-wall normal faults and an underlying detachment surface is best developed in the northern half of the Clifton quadrangle (Fig. 2-3). In contrast, the north-striking normal faults of set 3 cut the low-angle normal faults of set 1.

STRATIGRAPHY

Strata in the study area consist of Neoproterozoic to lower Paleozoic rocks unconformably overlain by the late Paleocene-early 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. 2-3).

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 in this area (Janecke, unpublished data) (Fig. 2-3). Within the Clifton Horst it unconformably overlies Ordovician strata in five different fault blocks and 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 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. 2-4). Pliocene-Pleistocene(?) gravel and conglomerate deposits are younger, newly identified strata 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. 2-2)) may be related to 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.

Synrift Deposits – Main Episode

The Salt Lake Formation in the Deep Creek half-graben consists of four members (Fig. 2-4) (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. 2-3). 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 near-shore lacustrine, fluvial, and deltaic deposits; and the overlying New Canyon Member; interpreted as a fluvial deposit (Figs. 2-3, 2-4) (Janecke and Evans, 1999). Altogether more than 2 km of Salt Lake Formation are preserved (Fig. 2-4).

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 conglomerates with tuffaceous, calcareous, and sandy groundmass and 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 consist mostly of lower Paleozoic carbonates with some chert and quartzite (Fig. 2-4). 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 of the Deep Creek half graben and on the west side of Oxford Ridge in the hanging wall of a folded low-angle normal fault (Fig. 2-3). It is at least 335 m thick in the Pocket Basin on the west edge of the Clifton Horst (Fig. 2-4). 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. 2-4). Bedding in the unit ranges from thick, massive, and fractured to thin and shaly. 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. The green coloring is most common and 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 Clifton Horst (Fig. 2-2), 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. 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 during deposition of the middle part of the member.

The presence of limestone beds with algal laminations at the base and top of the Cache Valley Member in the Deep Creek half graben (Fig. 2-4) 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 distal from the clastic wedge along the uplifted margins of the basin. Rare pebbly conglomerate beds near the top of the Cache Valley Member in the Deep Creek basin provide the only evidence for an emerging highland in the vicinity. We interpret the Valley Fault (Fig. 2-2) as the eastern margin of the basin at this time because closer normal faults such as the Clifton Cemetery Fault appear to have bounded the basin during deposition of overlying members. However, direct evidence for slip on the Valley Fault during deposition of the Cache Valley Member is lacking.

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. 2-3). It is >1.1 km thick east of the Clifton Horst and at least 1 km thick just west of the horst (Fig. 2-4). The Third Creek Member is also exposed in three places in the Clifton Horst in the south-central section of the Clifton quadrangle just west of Oxford Ridge (Fig. 2-3). 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 individual conglomerate beds range from 50 cm to 4 m thick. Groundmass of the conglomerate is variably tuffaceous, sandy, and calcareous. This unit also contains interbedded, slightly tuffaceous, and locally oolitic limestones with algal structures, shell fragments, molds of ostracod and gastropod shells, and chert nodules. Clasts of the conglomerate are generally subangular to well-rounded pebbles, and a few beds contain cobbles (Fig. 2-9a).

Sedimentary structures in the Third Creek Member suggest that it is a near-shore lacustrine and partly fluvial deposit (Fig. 2-9a). A 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. 2-3) with northeast-dipping topset beds and west-dipping foreset beds that indicate transport from east to west (Janecke and Evans, 1999). Limestone beds with ooids, molds of gastropod shells, 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. 2-10b).

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. 2-3). The unit is poorly consolidated in the study area. However, excellent consolidated exposures of this member west of Oxford Peak, northwest of the map area, consist of a parallel-bedded pebble to cobble conglomerates. The conglomerates are clast-supported, and clasts are uniformly well rounded with no fining- or coarsening-upward sequences (Fig. 2-9c, 2-9d, 2-9e). Clasts consist mostly of Brigham Group quartzites, Paleozoic carbonates and chert, and some recycled clasts from the Cache Valley Member of the Salt Lake Formation (Figs. 2-4, 2-11). The New Canyon Member is distinguished 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. 2- 4). Immediately 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 bedforms, and its persistence vertically and laterally. The New Canyon Member may represent overfilling of the basin by sediment.

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. 2-4, 2-11). Fifty to one hundred clasts >0.5 cm were counted within a 1 m^2 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 along the east side of 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. 2-4, 2-11). 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 average only 1% in conglomerates of the underlying Skyline Member.

Clasts of recycled Cache Valley Member average 26% of total clasts in the Third Creek Member, but make up only 3% of total clasts in the New Canyon Member (Fig. 2-4).

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. 2-4). 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. 2-4). Clasts counts 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. Zeolites are hydrated aluminosilicates of the alkalies and alkaline-earth minerals (Surdam, 1979). 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 successfully 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.

More detailed analyses of lithologically identical rocks in the Oneida Narrows area (Danzl, 1985) and other locations (Table 2-3) show that clinoptilolite is the principal zeolite species 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 setting, whereas the overlying Mink Creek Conglomerate was deposited by 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, or both.

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. 2-4, 2-9b, 2-11). 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. 2-9b). Unaltered equivalents of these siltstone and mudstone units of the Cache Valley Member are too poorly indurated to form coherent clasts able to survive appreciable transport, or to sustain or preserve fractures. XRD analyses confirm that the fractured clasts in the 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 2-3). 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 2-3). South of the Deep Creek area, our reconnaissance shows that zeolite-bearing zones interfinger with unaltered ash (point L on Fig. 2-2), and even farther south, zeolites are uncommon or absent (point J on Fig. 2-2). Zeolites reappear farther south at sites K, H and I (Table 2-3 and Fig. 2-2);
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 2-1). A common tectonic or climatic cause probably explains the synchronous development of saline/alkaline lakes in regions that are now 35 km apart (Fig. 2-10a);
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 mollusks (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.

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 2-1) (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 $^{40}Ar/^{39}Ar$ dating of an ash-flow tuff (1 sample) and chemical characterization and correlation of ash-fall tephra (21 samples). Broadly these age 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 2-1). Most of these ash-fall tuffs have the general compositional characteristics of those 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 2-1). 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), an unnamed ~8.4 Ma ash bed (sample rv89-9), the ~8.3 Ma Inkom ash bed, an 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 region, 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 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 the accumulation and deformation of the New Canyon Member predates 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. 2-9b) 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 portion 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., in press). 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 300m/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 2-2-2). 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 localized but 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 2-2) (Sacks and Platt, 1985; Goessel et al., 1999; Janecke and Evans, 1999).

Tuffaceous beds of the Cache Valley Member lap across 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 sub-basins of the larger Cache Valley Lake (Fig. 2-10a). Ash from the Yellowstone Hotspot filled the lake and settled quietly to the bottom. The initial slip on the Bannock detachment system is inferred to have begun at this time, but basin-margin facies of this Cache Valley Lake have not been identified. Creation of accommodation space outpaced sedimentation at this time (e.g., Carroll and Bohacs, 1999). The Cache Valley Member that is now preserved within and on either side of the Clifton Horst, was not influenced by any topographic highlands during deposition of the Cache Valley Member.

The following observations from the Salt Lake Formation across northern Cache Valley indicate that a single depositional basin probably existed during deposition of the Cache Valley Member of the Salt Lake Formation: 1) The Salt Lake Formation is exposed in widely separated fault blocks across Cache Valley but has consistent, distinctive lithologies vertically (Adamson et al., 1955; Danzl, 1982; 1985; Goessel et al., 1999; this study); 2) The Cache Valley Member always appears in the same stratigraphic position in these dispersed sections, and tephra-stratigraphic dating suggests that these similar rocks are about the same age; 3) The Cache Valley Member supplied significant detritus to the overlying Third Creek Member of the Salt Lake Formation in our study area and in the Cottonwood Valley/Oneida Narrows area (Danzl, 1985). We interpret this recycling as a record of uplift and tilting within the original large depositional basin of the Cache Valley Member; and 4) Some of the recycled clasts are angular (Fig. 2-9b), which suggests that the newly uplifted Cache Valley Member was proximal to these deposits.

During deposition of the Third Creek Member, the paleogeography of the rift basin changed dramatically from a single broad basin to numerous smaller east-northeast-tilted half graben and intervening tilt blocks (Fig. 2-10b). Uplifted rocks of the Cache Valley Member were stripped progressively from the emerging highlands and redeposited into these half graben. Through time larger highlands developed and deeper portions of the pre-Tertiary rocks were exposed (Fig. 2-10c). 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, 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. 2-10b). External drainage developed, and freshwater lakes replaced the saline/alkaline lakes of the prior episode.

Widespread subaerial conditions replaced the lacustrine, marginal-lacustrine to fluvial conditions during deposition of the New Canyon Member (Fig. 2-10c). 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. 2-4, 2-11). 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., in press).

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

Pliocene-Pleistocene(?) Piedmont 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. 2-3). This sequence is very thin compared 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. 2-7). 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. 2-3, 2-8). 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 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. 2-3, 2-7e, 2-8). 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. 2-7e). 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. 2-4, 2-11).

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 on the Dayton-Oxford Fault since deposition of these deposits than on the Deep Creek Fault. The steep rugged topography of the east side of the Clifton Horst supports this interpretation (Fig. 2-7a).

West of the Clifton Horst an erosionally dissected, mildly faulted and folded piedmont surface occupies the hanging wall of the Deep Creek Fault (Figs. 2-3, 2-7b, 2-7c, 2-7d). Small east- and west-dipping normal faults offset and folded the piedmont surface (Fig. 2-3). These faults strike northerly and are interpreted as synthetic and antithetic faults to the Deep Creek Fault (Figs. 2-5, 2-6).

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. 2-3). Just 1.2 km 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 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. 2-3). 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 gravel 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 southeast Idaho and northern Utah (Ore, 1982; Oaks et al., 1999; Ore, 1999; Kruger et al., in press). 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. 2-2). The Red Rock Pass area has well-developed gravel deposits that record north to south transport of boulders (DeVecchio, 2002). 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, in southernmost 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 valley-ward, 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 the 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 in Cache Valley show that there are only a few hundred meters of sediment preserved beneath the present valley floor of Cache Valley above the tuffaceous and tilted 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 east 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, because deposits are tilted, conformable with the Salt Lake Formation, and cut by a discordant erosion surface (Oaks, 2000 and Biek et al., 2001). Less accommodation space may have been 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 older than late Pleistocene 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. 2-2) (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; Oaks, 2000) 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 angular 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 2-2). 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 to 5.1 Ma (episode 3a, Table 2-2; Fig. 2-6). 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 2-2). 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. 2-3).

Renewed slip on the Dayton-Oxford normal fault elevated these alluvial-colluvial deposits at least 425 m above possibly correlative strata in northernmost Cache Valley and separated them from their downslope equivalents (episode 3c, Table 2-2). 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 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 to 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 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. 2-1). 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 2-2). 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 2-2; Fig. 2-10a). The broad basin was disrupted at least twice by normal faults (Table 2). The first structural “break-up” (episode 1c; Fig. 2-10) 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 margins of the modern horsts and graben (Fig. 2-10d). An incompletely understood episode of faulting along southeast-, east- and northeast-striking normal and normal-oblique slip faults (episode 2) apparently intervened between episodes 1 and 3 (Table 2-2).

Sedimentation occurred during episodes 1 and 3 of this evolution. Deposition of the Miocene-Pliocene Salt Lake Formation was associated with large-magnitude extension during episode 1, and records the transition from localized incipient rifting (episode 1a, Table 2-2) to an early coherent rift basin (episode 1b) to later structurally disrupted rift basins (episode 1c). Episode 1 was probably completed before 2 Ma.

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 external drainage to the Great Salt Lake sometime during episode 3 (e.g., Oaks et al., 1999) and might explain the sparse sedimentary record of this time throughout Cache Valley. Increases in accumulation rates interpreted from 3 of 4 oil wells in the Great Salt Lake between 3.1 and 2 Ma (Kowalewska and

Cohen, 1998) might record the integration of Cache Valley into the drainage basin of the Great Salt Lake, possibly in response to normal faulting in the Junction Hills area (Oaks, 2000).

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 highlands unlike 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. 2-1) (Janecke and Evans, 1999; Kruger et al., in press). The inferred listric breakaway, the Valley Fault, is the most continuous structure in the system and may persist 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., in press) in accordance with this interpretation (Fig. 2-1). 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; cf., Kruger et al., in press), 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. 2-2) (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 hypothesis and the alternative domino extensional style of 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. 2-2). 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. 2-2) (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 at least to the south end of Cache Valley (Evans and Oaks, 1996) (Fig. 2-2). 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 interpretation (Peterson and Oriel, 1970; Zoback, 1983). From Logan south, the Bannock detachment system has a steeper geometry. This change in geometry coincides with the southern extent of uplifted early Paleozoic rocks in the Cache-Pocatello Culmination (Fig. 2-1).

The amount of extension during episode 1 probably increased 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, 1999; Biek et al., 2001), the Malad Range (Janecke and Evans, 1999), Oxford Ridge (Link, 1982a, 1982b; this study) (Fig. 2-2), 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 north to Marsh Valley (Link, 1982a, 1982b; Goessel et al., 1999; Pope and Link, 2000; Biek et al., 2001; this study). Thus the greatest extension may be localized along a northeast transect through Oxford Ridge (Fig. 2-2) in the area where the Pocatello Formation is now exposed and the Deep Creek-Oneida Narrows saline/alkaline sub-basins developed during deposition of the Cache Valley Member. Analogous along-strike transitions of low-angle normal faults in highly extended terranes to moderately dipping normal faults in less extended areas have been documented in several parts of the Basin and Range province (Duebendorfer and Sharp, 1998; Miller et al., 1999; Faulds, 1999). Several segments of the East Cache Fault zone continued to be active normal faults during episode 3 (McCalpin, 1994).

The Cache-Pocatello Culmination is one of several culminations in the Sevier fold-and-thrust belt that localized large-magnitude extension. The Wasatch Culmination, south-southwest of Cache Valley (Fig. 2-1), also collapsed along normal faults, but extension began earlier, during deposition of the latest Eocene Norwood Tuff (Bryant, 1990; Constenius, 1996), and continues today along the Wasatch Fault and the East Great Salt Lake Fault (Yonkee, 1997; Mohapatra and Johnson, 1998). The much greater structural relief of the Wasatch Culmination (Schirmer, 1988; Yonkee, 1997) likely explains the earlier (late Eocene) onset of extension there relative to the later Miocene (~12 Ma) onset of extension in the Cache Valley area. Similarly, several culminations in the greater Salmon, Idaho, area collapsed to produce detachment faults in Paleogene time (Fig. 2-1) (Janecke et al., 2001). Thermo-tectonic factors might have triggered the extensional collapse but its location was probably controlled by the culminations. Tectonic models that relate the extension during episode 1 to passage of the Yellowstone Hotspot alone must be in error.

The similarities between the Bannock detachment system and the Sevier Desert Detachment are striking. Both systems collapsed structural culminations with fairly broad, flat-topped geometries (this study; DeCelles et al., 1995; Coogan and DeCelles, 1996). Both systems root to the west. Both formed in late Cenozoic time (Von Tish et al., 1985; Coogan and DeCelles, 1996; Stockli et al., 2001), although the Bannock detachment system initiated later and was short lived. Areas underlain by the shallow portions of both detachment faults expose large expanses of the Salt Lake Formation (Hintze, 1980). Both systems have breakaway faults ~150 km east of a metamorphic core complex produced by top-to-the east detachment faults. The Sevier Desert detachment is paired with the Snake Range Core Complex. The Bannock detachment system lies in a comparable antithetic structural position east of the Raft River-Grouse Creek-Albion Core Complex (Fig. 2-1). Perhaps patterns of crustal thickness in the thrust belt and its hinterland controlled the fundamental spacing of these extensional systems.

Breakup of the hanging wall of the Bannock detachment system (episode 1c) following initial extension (episode 1b) (Fig. 2-10) may be explained with the theoretical construct of an extensional wedge (Xaio et al., 1991; Fowler et al., 1995). This theory was developed by analogy with the critically tapered wedge theories that explain the behavior of contractional belts. An extensional wedge is bounded below by a low-angle normal fault and may break up along internal normal faults if the taper of the wedge is too great, if the strength of the low-angle fault increases, if the wedge weakens internally (possibly due to hydrothermal alteration), if sedimentation changes the taper of the wedge, or if isostatic rebound of the footwall changes the configuration of the wedge (Xiao et al., 1991; Fowler et al., 1995). A broad north-northwest-trending anticline is defined by foliation in the Pocatello Formation is mapped in the footwall of the Bannock detachment system on Oxford Ridge (Figs. 2-3, 2-5, 2-6) and suggests that footwall uplift probably played a role in the break-up of the broad Cache Valley lake basin during episode 1c (Fig. 2-10c).

CONCLUSIONS

Geologic studies in the 7.5-minute Clifton and Malad City East quadrangles provide 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 spaced normal faults in their hanging walls 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, shows 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 southern 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 during this time and 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, northwest 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 spaced 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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Table 2-1. Tephrochronologic age determinations in the Salt Lake Formation of south-central Idaho.

Sample ID	Member	Location (Township and Range /Lat and Long/UTM)	Ash Bed type* Ash Bed	Age Range or Age (Ma)**	Comments (stratigraphic position based on contacts and structure)
jce156-98	Skyline (Ts)	SE/4, SW/4, SW/4, Sec 8, T15S, R37E (42° 07' 41" N, 112° 09' 48" W) $^{46}64^{600}$ N, $^{4}03^{800}$ E	Twin Falls "d2"	10.5-8.0[2]	Stratigraphic position is likely high in section within Ts
jce187-98	Cache Valley (Tcv)	SE/4, NE/4, SE/4, Sec 31, T14S, R37E (42° 09' 34" N, 112° 10' 06" W) $^{46}68^{200}$ N, $^{4}03^{400}$ E	Twin Falls Wooden Shoe Butte?	10.5-8.0[2] 10.13 ± 0.03[1]	Stratigraphic position is likely relatively high within Tcv
jce178-98	Cache Valley	NE/4. NW/4, NE/4, Sec 19, T14S, R37E (42° 11' 51" N, 112° 10' 24" W) $^{46}72^{400}$ N, $^{4}03^{200}$ E	Twin Falls Wooden Shoe Butte?	10.5-8.0[2] 10.13 ± 0.03[1]	Stratigraphic position is high within Tcv
jce122-98	Cache Valley	SW/4, SE/4, SE/4, Sec 36, T13S, R36E (42° 14' 33" N, 112° 11' 34" W) $^{46}77^{400}$ N, $^{4}01^{700}$ E	Twin Falls Wooden Shoe Butte?	10.5-8.0[2] 10.13 ± 0.03[1]	Stratigraphic position is very high within Tcv
suj303-98	Cache Valley	NE/4, NE/4, Sec 19, T14S, R37E (42° 11' 47" N, 112° 10' 23" W) $^{46}72^{037}$ N, $^{04}03^{170}$E	Twin Falls Wooden Shoe Butte?	10.5-8.0[2] 10.13 ± 0.03[1]	Stratigraphic position is likely high within Tcv
jce137-98	Third Creek	SW/4, NE/4, SE/4, Sec 20, T14S, R37E (42° 11' 22" N, 112° 09' 00" W) $^{46}71^{500}$ N, $^{4}05^{100}$ E	Twin Falls? unique	10.5-8.0[2]	Stratigraphic position is low within Ttc
jce181-98	Third Creek	SW/4, SW/4, SW/4, Sec 31, T13S, R37E (42° 14' 35" N, 112° 11' 22" W) $^{46}77^{500}$ N, $^{4}01^{800}$ E	Twin Falls "a"	10.5-8.0[2]	Stratigraphic position is likely very low within Ttc
jce202-99	Third Creek (Ttc)	SE/4, NW/4, SE/4, Sec 17, T14S, R37E (42° 12' 11" N, 112° 09' 10" W) $^{46}73^{000}$ N, $^{4}04^{900}$ E	Twin Falls "a"	10.5-8.0[2]	Stratigraphic position is approximately mid-Ttc (this sample correlates with 39-97)
jce53-97	Third Creek	NE/4, SW/4, SW/4, Sec 14, T14S, R37E (42° 12' 07" N, 112° 06' 17" W) $^{46}72^{400}$ N, $^{4}08^{800}$ E	Twin Falls "b" (?)	10.5-8.0[2]	Stratigraphic position is unknown within Ttc
jce50-97	Third Creek	NE/4, SW/4, NE/4, Sec 23, T14S, R37E (42° 11' 37" N, 112° 05' 45" W) $^{46}79^{600}$ N, $^{4}09^{500}$ E	Twin Falls "b"	10.5-8.0[2]	Stratigraphic position is unknown within Ttc
smc58	Third Creek	NE/4, NE/4, SW/4, Sec 10, T15S, R37E $^{46}64^{000}$N $^{4}07^{500}$E	Twin Falls	10.5-8.0[2]	Southwest side of Clifton Horst
smc33	Third Creek	SE/4, SE/4, NW/4, Sec 7, T15S, R38E $^{46}64^{200}$N $^{4}13^{600}$E	Twin Falls (rv88-4)	10.5-8.0[2] ~9.6 ± 0.2[3]	Ash interbedded with lacustrine limestones within the Clifton Horst in hanging wall of the Clifton detachment fault
jce121-98	Third Creek	NW/4, SE/4, NE/4, Sec 28, T14S, R37E (42° 10' 51" N, 112° 07' 53" W) $^{46}70^{200}$ N, $^{4}06^{300}$ E	Twin Falls Inkom	10.5-8.0[2] ~8.3 ± 0.5[2]	Stratigraphic position is unknown within Ttc

jce166-98	Third Creek	SE/4, NE/4, SE/4, Sec 9, T14S, R37E (42° 13' 02" N, 112° 07' 48" W) $^{46}74^{600}$ N, $^{4}06^{700}$ E	Twin Falls “a1”	10.5-8.0[2] ~8.1 ± 0.2[2]	Stratigraphic position is very high within Ttc (This age conflicts with age of sample jce209-99)
smc67	Third Creek	SE/4, SW/4, SE/4, Sec 4, T15S, R38E $^{46}65^{800}$N $^{4}16^{000}$E	Twin Falls “a1”	10.5-8.0[2] ~8.1 ± 0.2[3]	East of Clifton Horst in the middle of the Third Creek Member
smc68	Third Creek	SE/4, SW/4, SW/4, Sec 4, T15S, R38E $^{46}65^{500}$N $^{4}15^{800}$E	Twin Falls “a1”	10.5-8.0 ~8.1 ± 0.2[3]	East of Clifton Horst in the middle of the Third Creek Member
jce58-97	Third Creek	NE/4, NW/4, SW/4, Sec 14, T14S, R37E (42° 10' 34" N, 112° 07' 30" W) $^{46}72^{600}$ N, $^{4}08^{800}$ E	Twin Falls Rush Valley	10.5-8.0[2] ~7.9 ± 0.5	Stratigraphic position is unknown within Ttc
smc55	Third Creek	SW/4, SE/4,SE/4, Sec 34, T14S, R38E $^{46}67^{400}$N $^{4}17^{000}$E	Twin Falls Rush Valley	10.5-8.0[2] ~7.9 ± 0.5	East of Clifton Horst
jce204-99	Third Creek	NW/4, NE/4, SW/4, Sec 17, T14S, R37E (42° 12' 19" N, 112° 09' 36" W) $^{46}73^{200}$ N, $^{4}04^{200}$ E	Twin Falls? rv89-9	 8.4 ± 0.5[3]	Stratigraphic position is approximately mid-Ttc
jce209-99	Third Creek	NE/4, SW/4, NE/4, Sec 9, T14S, R37E (42° 13' 22" N, 112° 10' 28" W) $^{46}75^{300}$ N, $^{4}06^{300}$ E	Heise Santee	7.5-4.0[2] 5.1 ± 0.1[2]	Stratigraphic position is very high within Ttc (This age conflicts with age of sample jce166-98)
jce207-99	Third Creek	SE/4, SW/4, NW/4, Sec 21, T14S, R37E (42° 11' 35" N, 112° 08' 42" W) $^{46}72^{800}$ N, $^{4}05^{500}$ E	Heise Santee/Kilgore	7.5-4.0[2] 5.1-4.4[2]	Stratigraphic position is likely high within Ttc

*Ash Bed type – All ash beds lie in the compositional field of Yellowstone hotspot ash-fall tuffs. In the Salt Lake Fm. most have the general compositional characteristics of either ash beds from sources in the Twin Falls volcanic field (10.5-8.0 Ma) or the Heise volcanic field. Twin Falls-type ash beds are often compositionally heterogeneous and difficult to correlate with confidence.

Ash Bed – Name or sample is of likely correlative ash bed. Ash bed names/id’s from Perkins et al. (1998), Perkins and Nash (2002), or this study (“a”, “a1”, “b”, “d2”).

**Data sources: (1) Perkins et al. (1998); (2) Perkins and Nash (2002); (3) this study.

Table 2-2. Summary of tectonic events in the northern Cache Valley area.

Designation	Structural developments	Sedimentary deposit or unconformity associated with this event	Age (based on table 1)	Location
Episode 1 Fault set 1a	Small localized rift basins filled with mostly conglomerate	Skyline Member of the Salt Lake Formation, Collinston Conglomerate, and Red Conglomerate	Older than 10.27±0.07 Ma ($^{40}Ar/^{39}Ar$ date) and probably between 12 and 10.3 Ma	Western Deep Creek Basin, Cottonwood Valley (Sacks and Platt, 1985) and northern Wellsville Mountain (Goessel et al., 1999)
Episode 1 Fault set 1b	Initiation of the Bannock detachment system and gravitational collapse of Cache-Pocatello culmination; development of Clifton-Oxford anticline	Cache Valley Member of Salt Lake Formation	Before 10.27±0.07 Ma until about 10.13± 0.3 Ma	The entire area of Fig. 10a
Episode 1 Fault set 1c	Breakup of the hanging wall of the Bannock detachment system along SW-dipping normal faults; development of Clifton basin anticline	Third Creek and New Canyon Members	Before 9.6 ± 0.2 Ma until after 5.1 to 4.4 Ma (possibly until about 2 to 3 Ma)	Cottonwood Valley/Oneida Narrows area westward at least to Deep Creek half graben
Episode 1 Fault set 1d	Development of Oxford Ridge anticline; Clifton fault steps to present position, excising its hanging wall.			
Episode 2 Fault set 2 (Hot spot-related flexure and faulting)	Formation of cross faults with E and NE strikes	Little or no sedimentary record?	After episode 1c and before episode 3a; after 5.1 to 4.4 Ma	In broad zone adjacent to the Eastern Snake River Plain (Janecke et al., 2000)
Episode 3a Fault set 3 (small-magnitude extension)	Modern Basin and Range topography dismembers the older rift basins along N-striking normal faults; Clifton Horst, Malad Range, and Clarkston Mountain are uplifted. Development of Pocket basin syncline?	Erosion and removal of most of the Salt Lake Formation, Wasatch Formation, lower Paleozoic and Neoproterozoic rocks from the Clifton Horst	After 5.1 to 4.4 Ma	Clifton Horst, Malad Range, Clarkston Mountain
Episode 3b Fault set 3	Tectonic quiescence and near burial of the Clifton Horst by piedmont gravel deposits	Deposition of piedmont gravel	After 5.1 to 4.4 Ma Pliocene-Pleistocene(?)	Figs. 3, 10D Deep Creek half graben, remnants on Clifton Horst
Episode 3c Fault set 3	Renewed uplift of the Clifton Horst	Erosion of piedmont gravel; external drainage to the Great Salt Lake was established by this time, thus lowering the base level	After deposition of piedmont gravel; after Early to mid Pleistocene	Clifton Horst, Malad Range, Clarkston Mountain

Table 2-3. Locations and attributes of zeolite occurrences in the Salt Lake Formation in the Cache Valley area. See Figure 2 for locations of samples.

Location	Quadrangle	Thickness of zeolite- bearing zone	Locations of unaltered tuffs relative to the zeolite-bearing zones	Source of data	Minerals, if known, and method of detection
A. Deep Creek half graben	Malad City East, Idaho	Thick, about 400-600 m	Rare unaltered tephras interfinger with the zeolite-bearing beds near the top of the section	J. C. Evans and Janecke, unpublished data	
B. Pocket Basin	Clifton, Idaho	381 m	Pervasive alteration	This study	
C. Clifton Horst (excluding Pocket basin)	Clifton, Idaho	Variable, at least tens of m	Interfingering of zeolite-bearing beds with algal limestones and thick unaltered tephras	This study	Clinoptilolite, XRD
D. Gravel pit east of Clifton Horst at Five Mile Canyon	Weston Canyon, Idaho	Thin, > 10 m	Relationship to tephras and limestone-bearing beds is uncertain	Janecke, unpublished mapping	
E. Northern end of Oxford Ridge and environs	Scattered localities in Oxford and Malad Summit, Idaho	unknown	No data	Link, 1982a	
F. Cottonwood Valley	Cottonwood Peak, Swan Lake, Treasureton, Idaho	Medium thickness?	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)	Sacks and Platt (1985); Sacks (1984)	Clinoptilolite, visual identification of concretions (p. 40 in Sacks and Platt, 1985)
G. Oneida Narrows	Riverdale, Idaho	380 m	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	Danzl (1985)	Clinoptilolite, authigenic silica, montmorillonite, calcite, illite: XRD, petrography, SEM
H. Amoco Lynn Reese well	Smithfield, Utah	36 m	Unaltered tuffs underlie and overlie the zeolite-bearing strata	Brummer (1991)	Clinoptilolite, XRD
I. Junction Hills	Cutler Dam, Clarkston, Utah	98 m	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 zeolite beds	Goessel (1999); Oaks (2000)	
J. Steel Canyon area Note: This is North Canyon of Adamson et al. (1955)	Henderson Creek, Idaho, and northernmost Clarkston Mountain, Utah	No zeolites	No data	Janecke; Biek; and Perkins, unpublished reconnaissance	
K. Newton Mountain area	Trenton, Utah	Unknown	No data	J. Boettinger and P. Kolesar, oral comm.	Clinoptilolite, XRD
L. Rattlesnake Ridge area, south of Five Mile Canyon	Weston Canyon, Idaho	Very thick, unless repeated by faulting	Intertongued tuffaceous beds	R.Q. Oaks, Jr., data from sites for gravity survey, 2000-2001; Janecke	

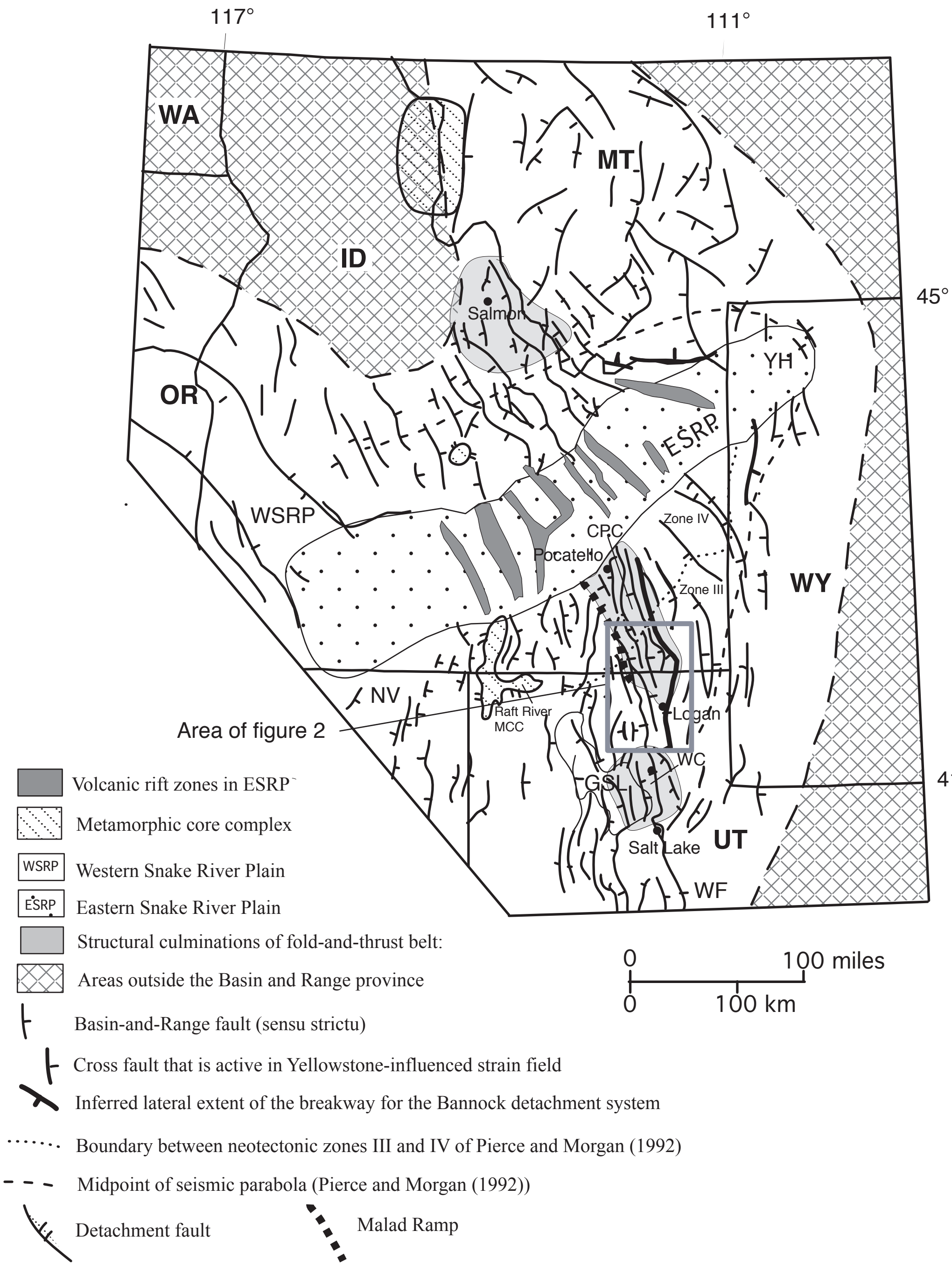


Figure 2-1. Regional map showing the major active normal faults adjacent to the Eastern Snake River Plain (ESRP), the north-northwest-trending Cache-Pocatello culmination (CPC), and selected other features. Note that the inferred Bannock detachment system essentially coinsides with the Cache-Pocatello culmination, and has the opposite sense of vergence as the partly coeval Raft River metamorphic core complex. The areally 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.

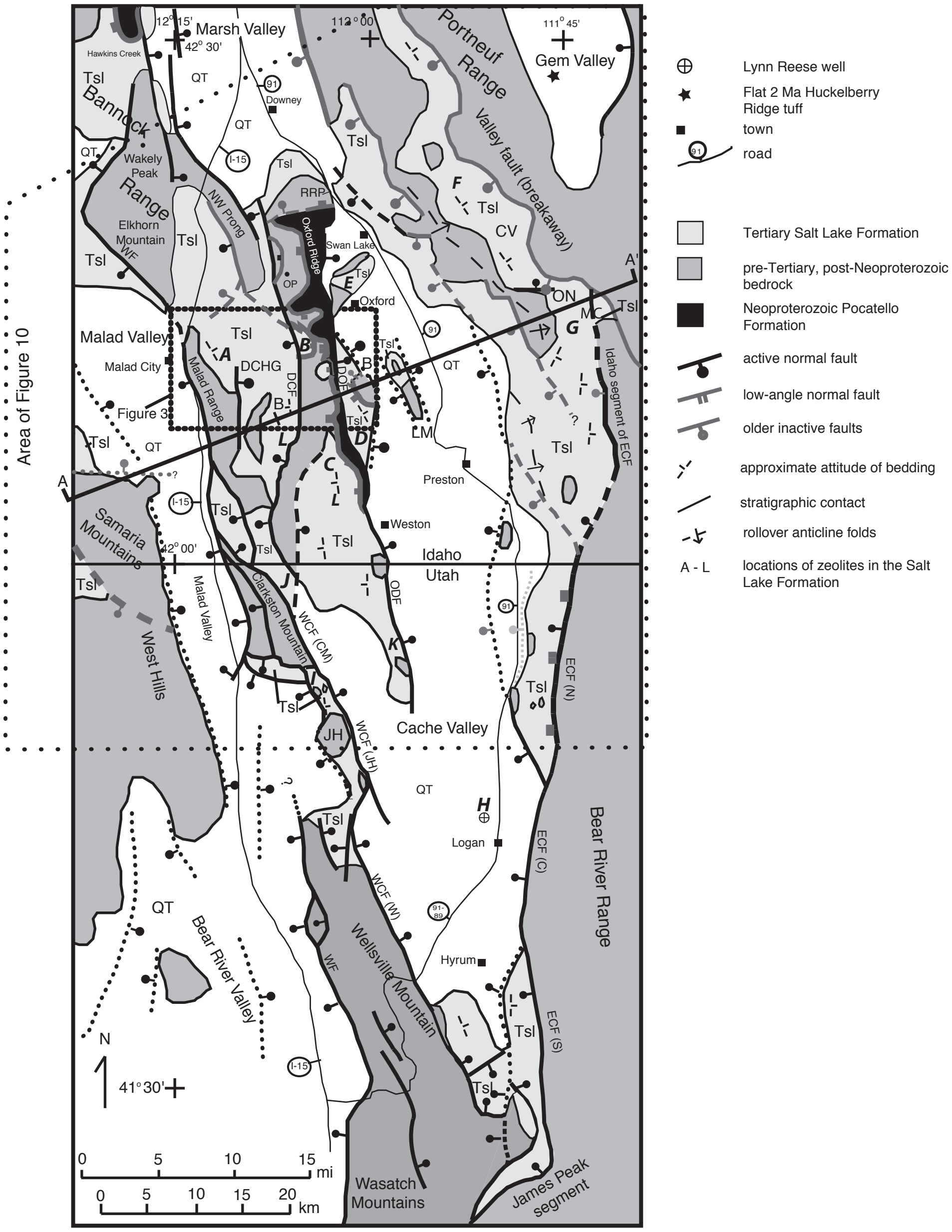


Figure 2-2. Simplified geologic map showing the distribution of the Salt Lake Formation around Cache Valley, Idaho and Utah, and the active Basin and Range normal faults. Some older normal faults are also shown(grayed). The Clifton horst is the up-thrown block between the Deep Creek and Dayton-Oxford 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 segments), WF = Wasatch fault. Some buried faults (dotted) from Zoback (1983). Letters A-L refer to entries in Table 2-3. A-A' and B-B' are geologic cross sections in Figures 2-5 and 2-6, respectively. This map supercedes a compilation in Janecke and Evans (1999) and includes data from Goessel et al. (1999) and Oaks et al. (1999).

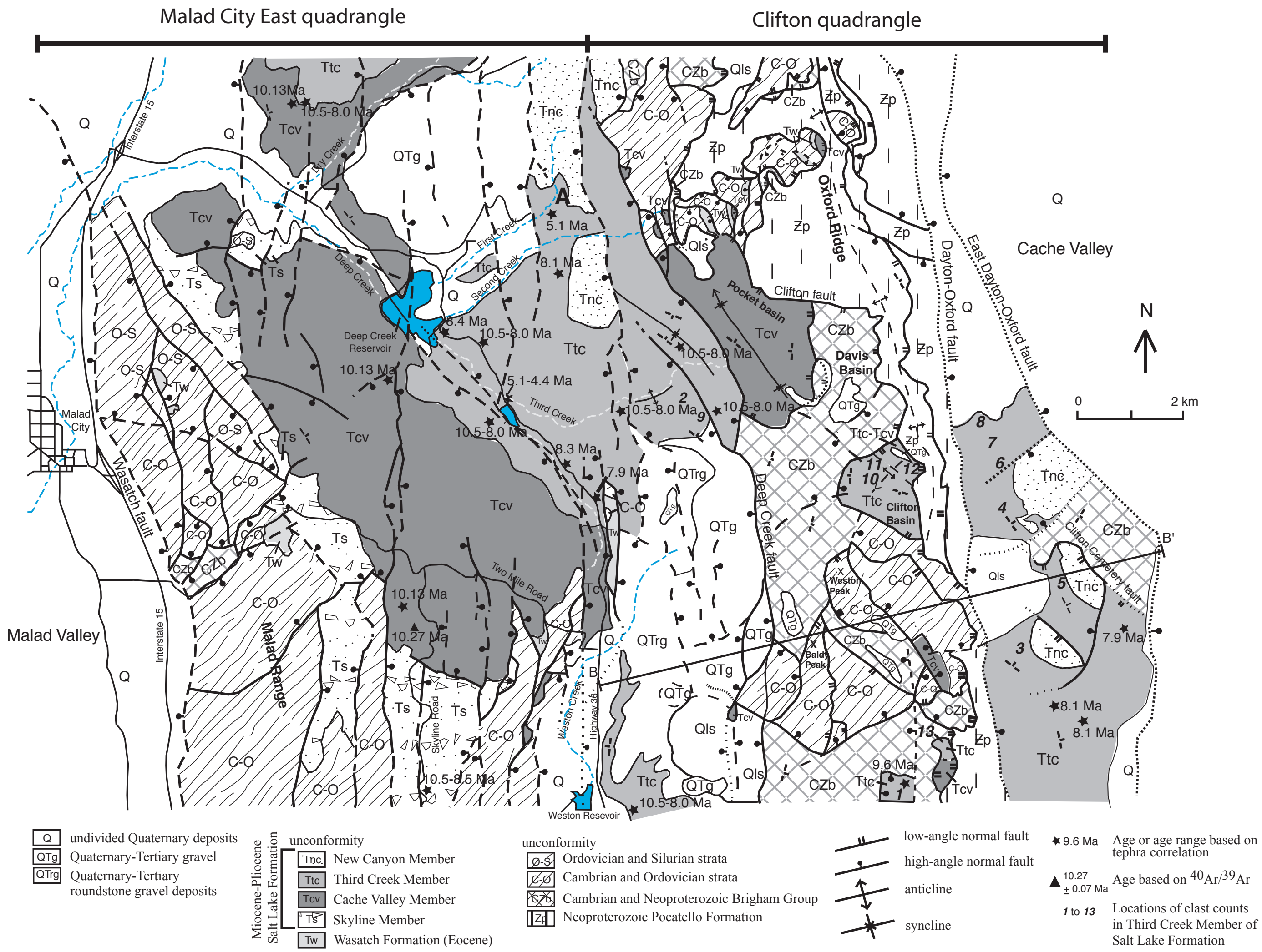


Figure 2-3. Simplified geologic map of the Clifton and Malad City East 7.5-minute quadrangles. See Figure 2-2 for location. The contact between Cache Valley and Third Creek members may be angular in some places.

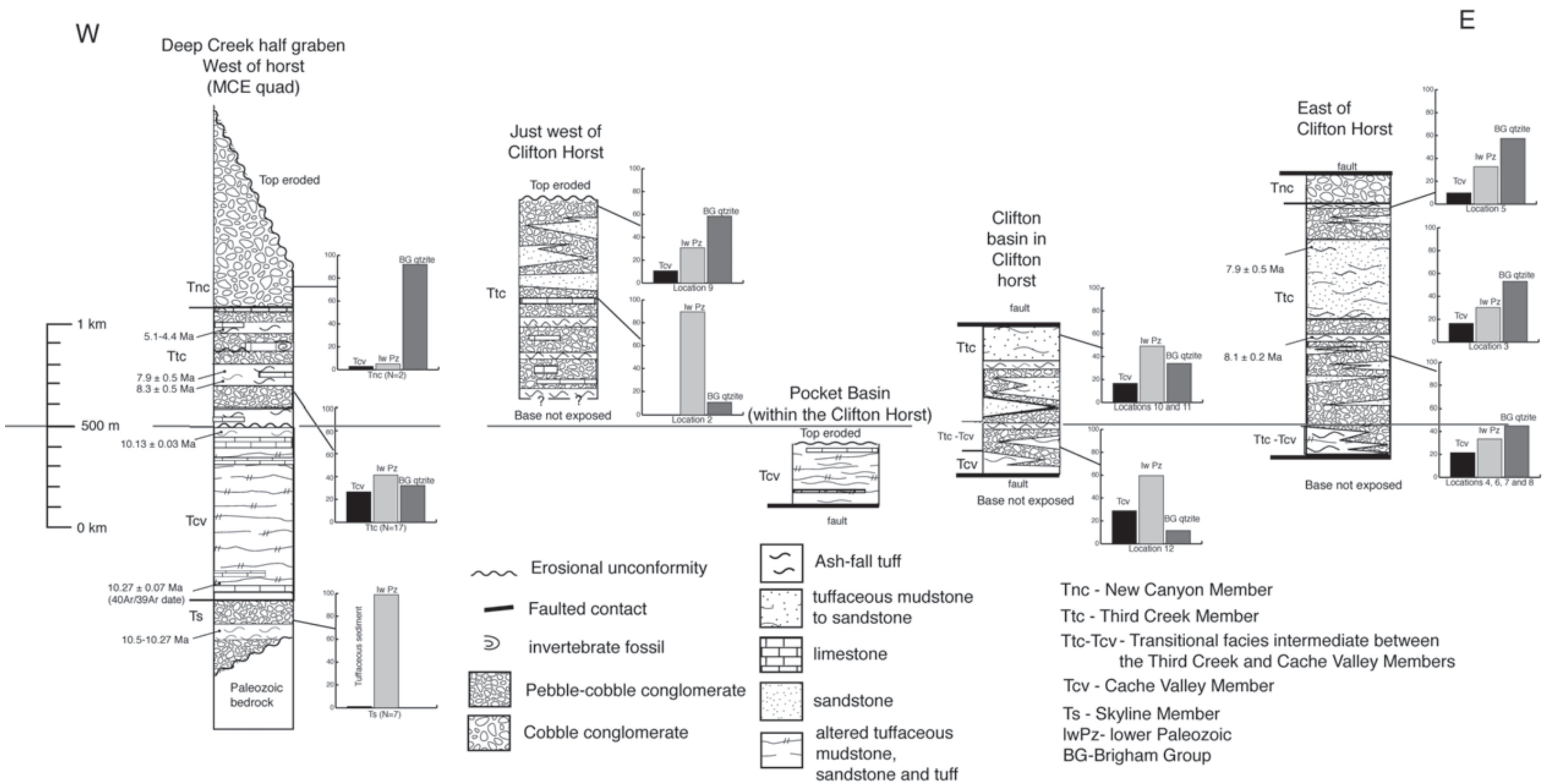


Figure 2-4. Schematic stratigraphic columns and clast counts of the Salt Lake Formation across the Clifton Horst and in the Deep Creek half graben. Note that clast counts from east of the horst show an unroofing seqence of the source area(s) vertically upward within the Third Creek Member, whereas clast 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.

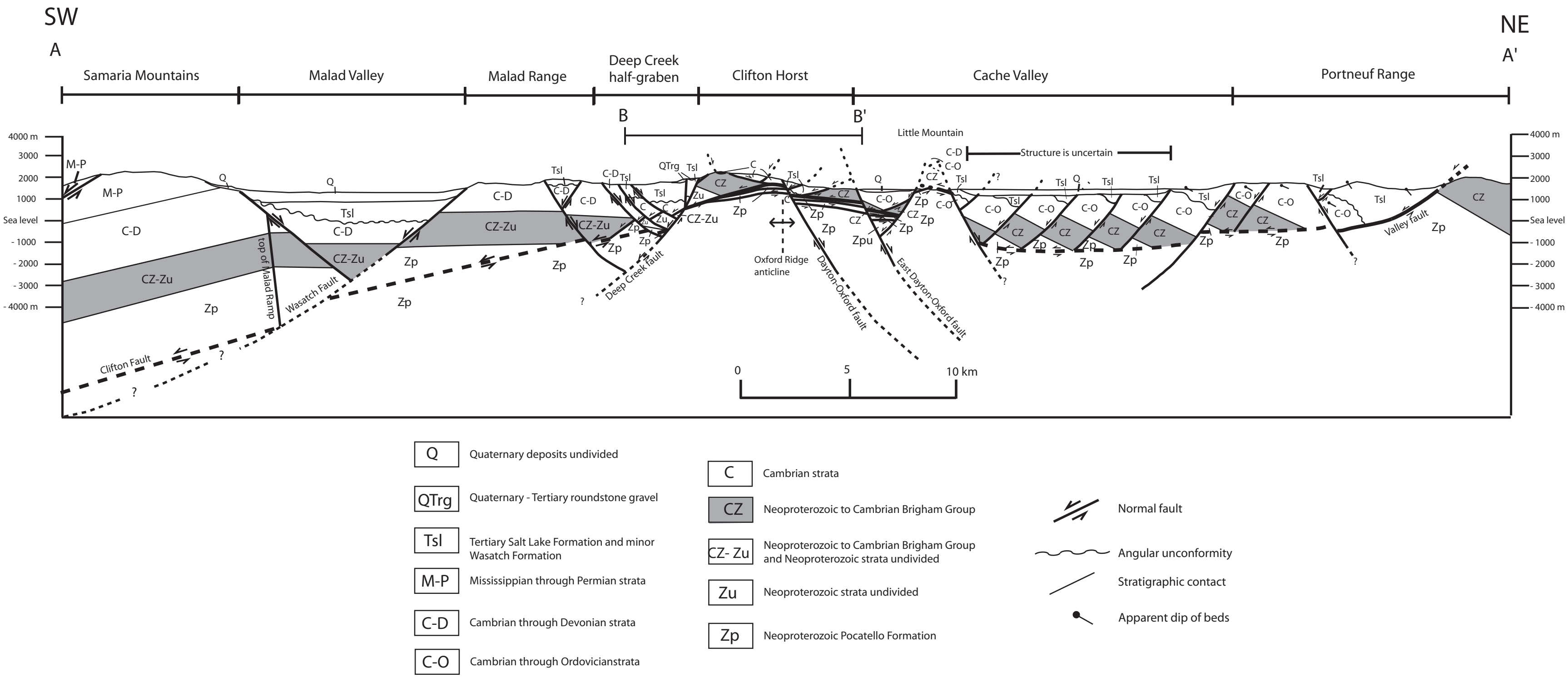


Figure 2-5. Regional geologic cross section, A-A', of the northern Cache Valley, Idaho area. The structure beneath Cache Valley is uncertain. Surface geology was compiled from Wach (1967), Platt (1977), Oriel and Platt (1968, 1980), Link and LeFebre (1983), Danzl (1985), Janecke and Evans (1999), Janecke et al. (in press), Janecke, unpublished mapping, and this study. Peterson and Oriel (1970) provided some subsurface control. Restoration indicated ~60% extension. See Figure 2-2 for location of this section and section B-B'. Approximate position of the Malad Ramp (Rodgers and Janecke, 1992) is shown.

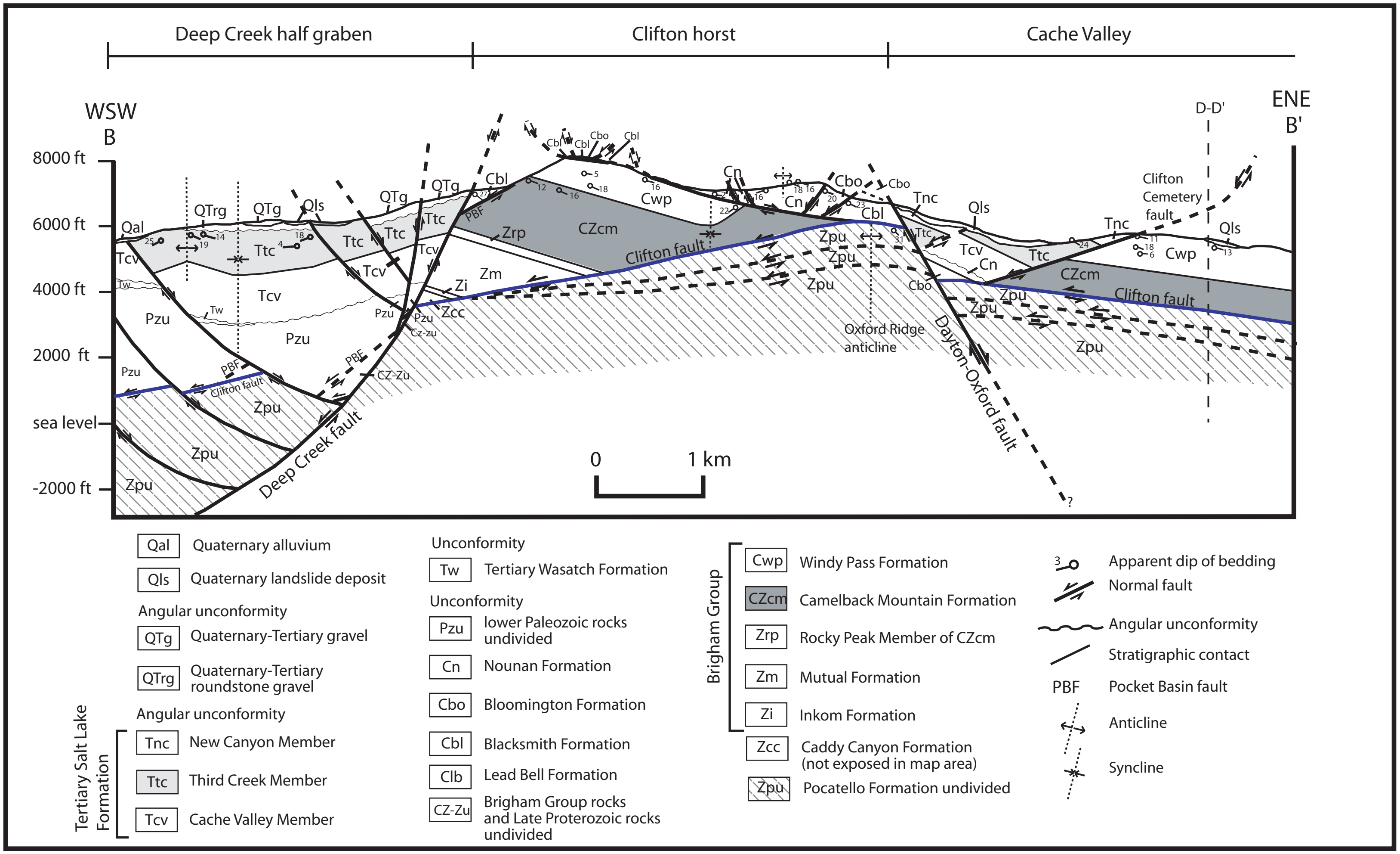
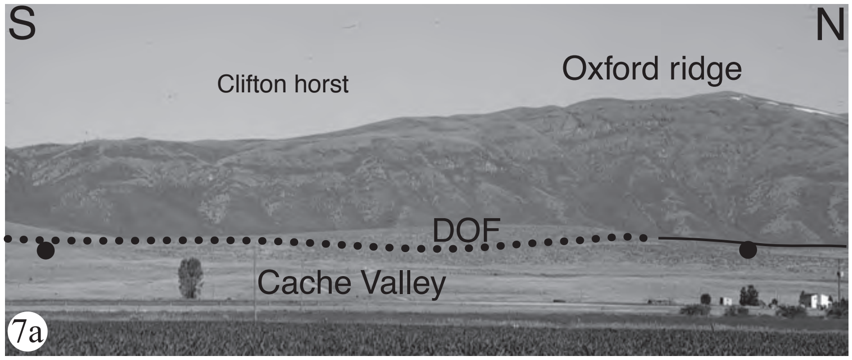


Figure 2-6. Geologic cross section B-B' across the Clifton horst in the study area. See Figure 2-3 for location of this section. Relationships beneath Clifton fault are uncertain. The Clifton fault is the main structure of the Bannock detachment system in this area. Probable angular unconformity between Cache Valley and Third Creek members of the Salt Lake Formation is not shown because its geometry is not well characterized.



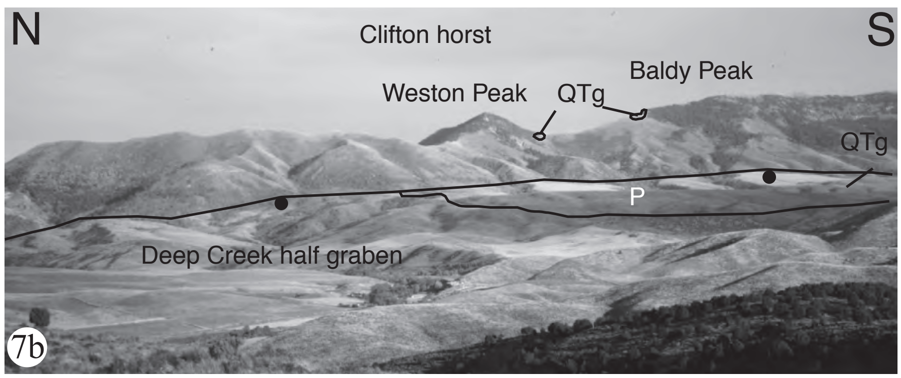


Figure 2-7. Photographs illustrating the geomorphology of the Clifton Horst and the associated Pleistocene (?)-Pliocene piedmont gravel deposits. (a) Steep east face of the Clifton Horst. No piedmont gravel depsits are preserved on Oxford Ridge or in Cache Valley. Note the topographic contrast between the east and west sides (Figs. 2-7b, c, d) 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 deposists 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 2-8 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 2-3). Bedding attitudes parallel the piedmont surfaces in 2-7c and 2-7d.

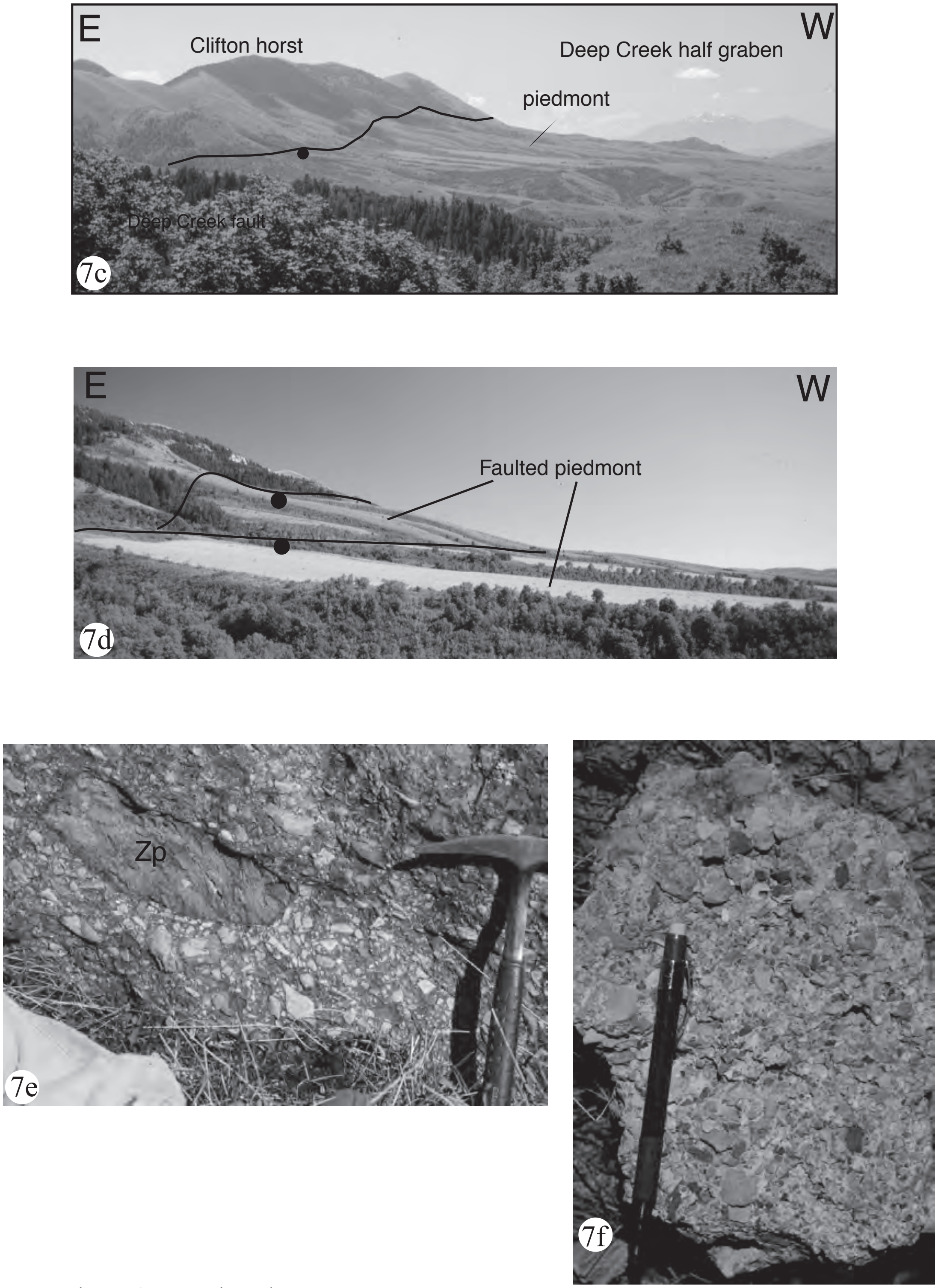


Figure 2-7 continued.

Figure 2-8. The Clifton low-angle normal fault cuts tilted Salt Lake Formation (with dashed traces of bedding) subsequently overlain by piedmont gravel in Clifton basin. Relations immediately west of the photograph are included in a sketch. Abbreviations are: QTg = Pliocene-Pleistocene(?) piedmont gravel deposits; Ttc/Tcv = transitional unit between the Cache Valley and Third Creek Members of the Salt Lake Formation; Zps = Scout Mountain Member of the Pocatello Formation. Figure 2-7e is a close-up of the piedmont deposits in this area.

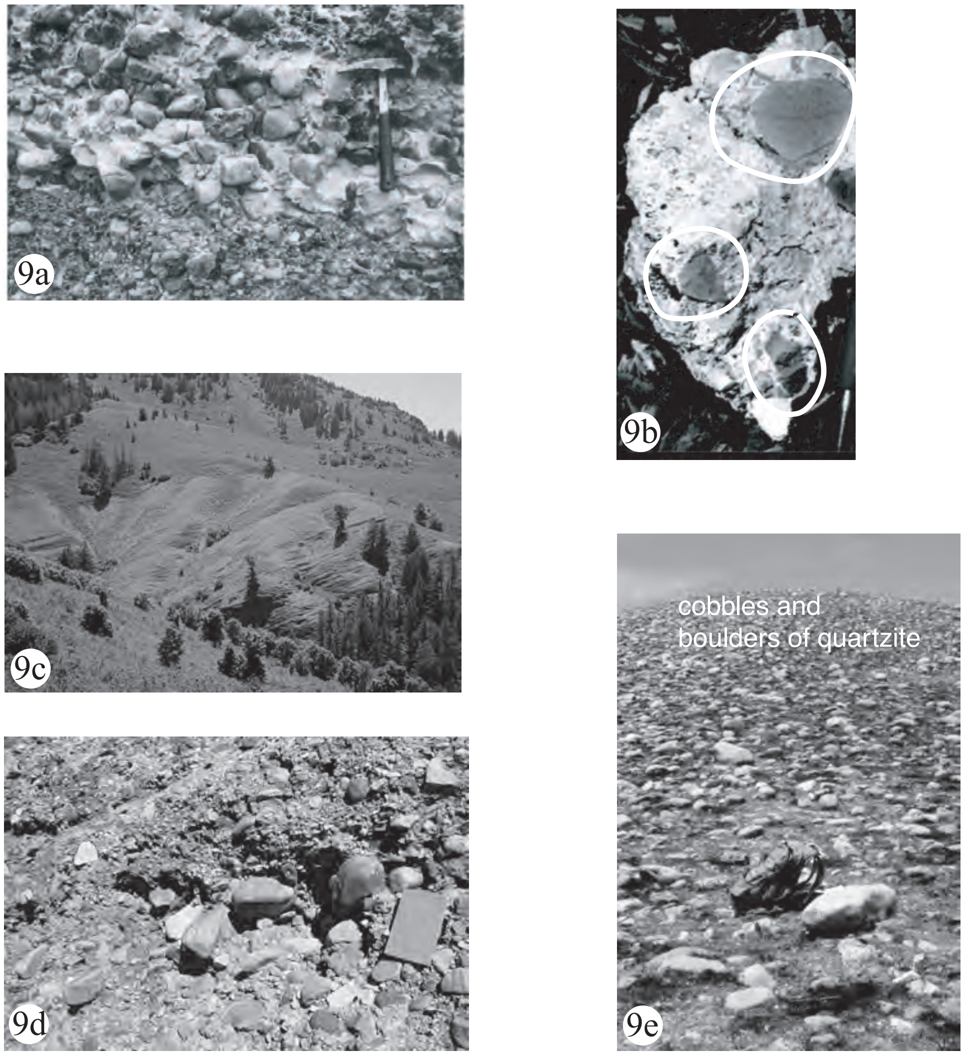


Figure 2-9. Photographs of important conglomeratic units in the Salt Lake Formation. (a) Sub-rounded to rounded clasts of predominantly Brigham Group quartzites in conglomerate beds of the uppermost Third Creek Member of the Salt Lake Formation, west of Clifton horst. Note the rather good sorting of 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. (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.

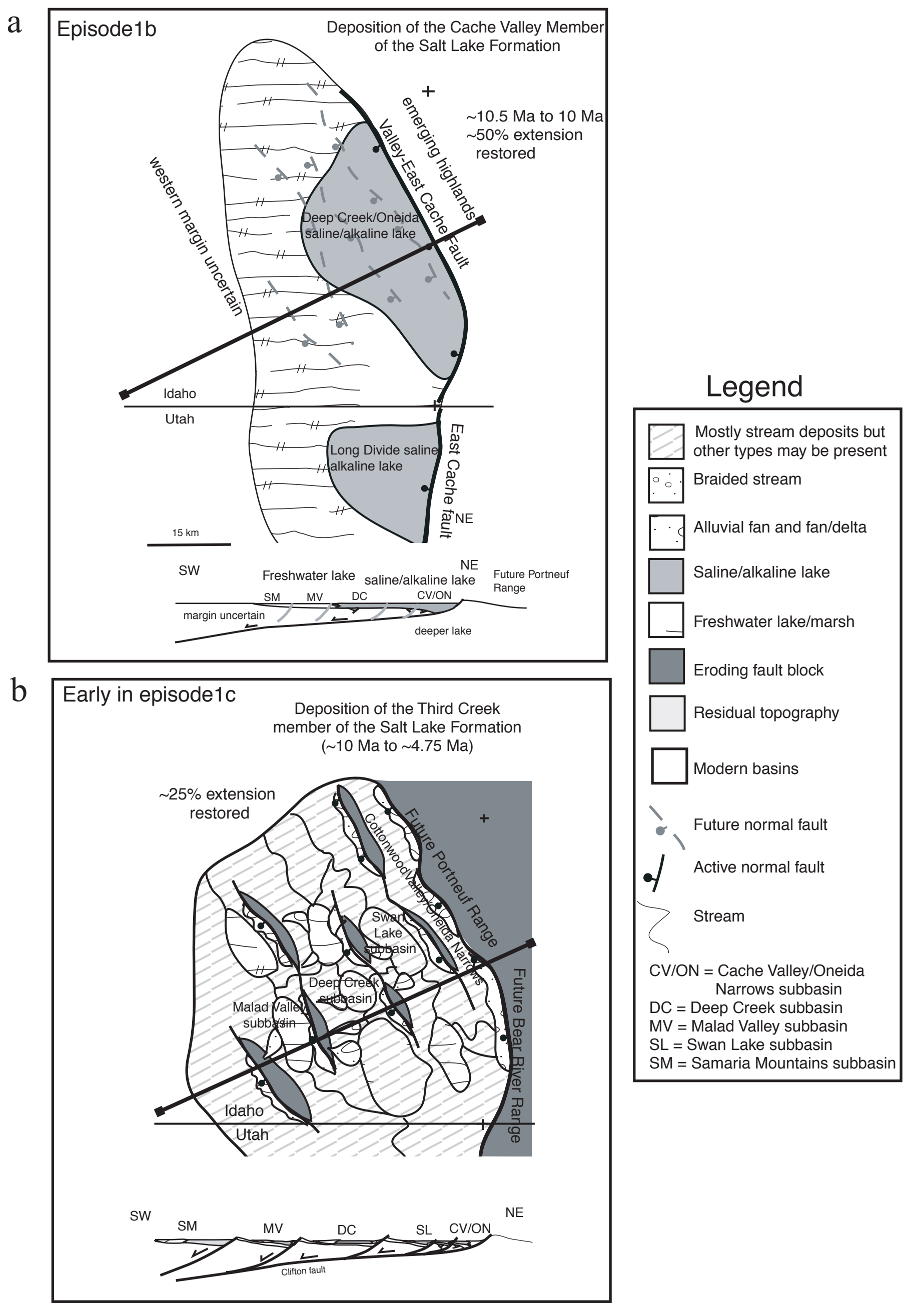


Figure 2-10. Reconstructed paleogeographic maps and cross sections of the northern Cache Valley area between the modern West Hills/Samaria Mountains and the Bear River Range. Extension was reconstructed based on a preliminary reconstruction of Figure 2-5. Episodes of extension correspond to those in Table 2-2. 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.

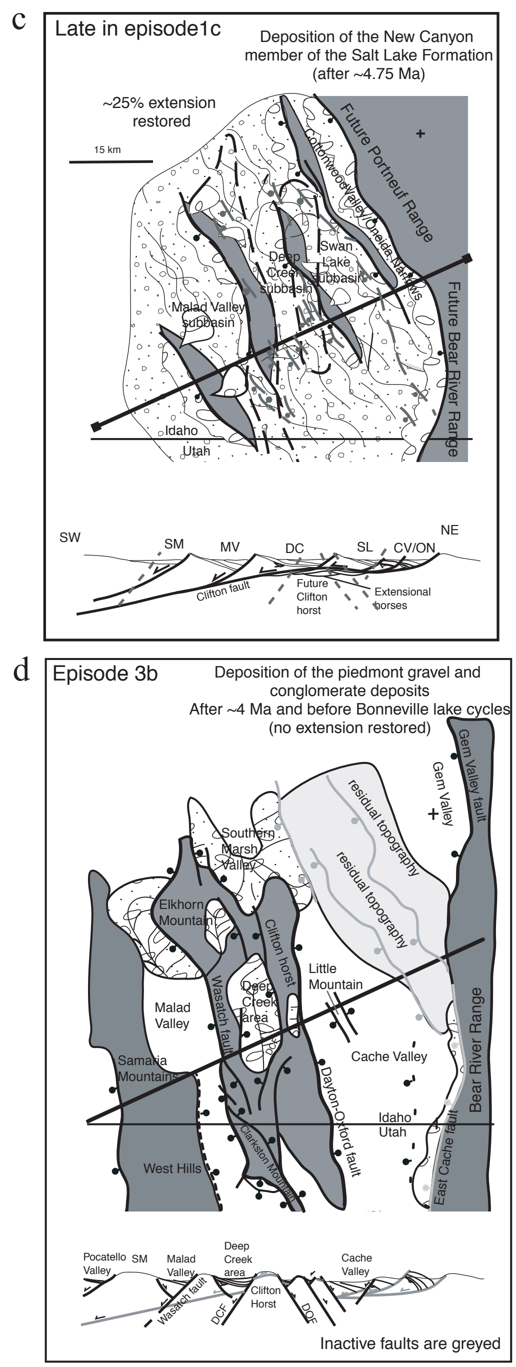


Figure 2-10 cont. (c) Late during episode 1c. Exhumation of tilted blocks exposed more pre-Tertiary bedrock. Braided streams filled the Deep Creek sub-basin. External drainage. The future positions of the Clifton Horst and the Elkhorn Mountain-Clarkston Mountain fault blocks are outlined with dashed lines. (d) Slip on southwest-dipping Bannock detachment system ceased. High-angle normal fault offset initiated along north-striking Basin-and-Range faults; 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.

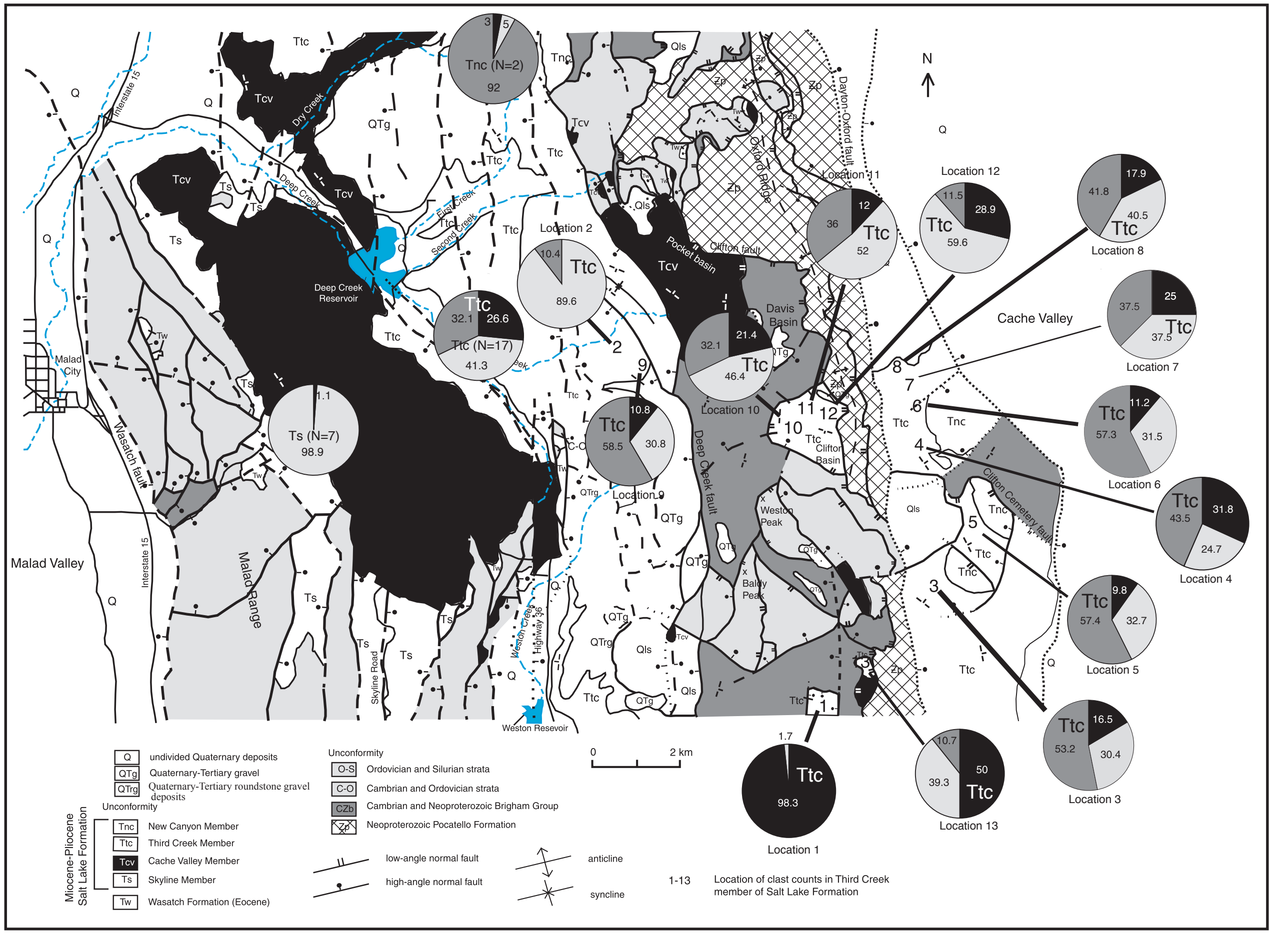


Figure 2-11. Compositions of conglomerates in the Salt Lake Formation superimposed on a simplified geologic map. The map depicts lateral changes in composition within the Third Creek Member (all but two locations) and shows the positions of conglomerates relative to potential source areas in adjacent highlands. Vertical changes in composition are shown in Figure 2-4. Note the similar compositions of the Third Creek Member east of, west of, and within the Clifton Horst. Compositions are variable within the Third Creek Member, but the overall assemblage is distinctive and is easy to discriminate from conglomerates of the Skyline Member and New Canyon Member.

CHAPTER 3

STRUCTURE AND EVOLUTION OF A LOW-ANGLE NORMAL-FAULT SYSTEM IN THE SOUTHERN BANNOCK RANGE, SOUTHEAST IDAHO[1]

ABSTRACT

Geologic mapping of the Clifton 7.5-minute quadrangle indicates that the bulk of Cenozoic extension in southeast Idaho was accommodated by slip on a low-angle normal-fault system, the Bannock detachment system. The Bannock detachment system is a regionally extensive system of low-angle normal faults with a possible N-S extent of >130 km. The detachment system was active from ~12 Ma to 4 Ma (Janecke et al., in press; Chapter 2 of this thesis) and accommodated >15 km of WSW-directed extension. The footwall is composed of Neoproterozoic Pocatello Formation and Tertiary intrusions that were metamorphosed to greenschist and amphibolite facies. A mafic sill invaded along a low-angle fault of the Bannock system during extensional exhumation and contains both ductile and brittle fabrics. Cross-cutting relationships show that the master detachment fault, the Clifton fault, is the youngest low-angle normal fault of the system, was active at a low angle, and has not been rotated to a low-dip angle through time. Map patterns and relationships indicate that the hanging wall to the detachment system began as a cohesive block that later broke up along listric and planar normal faults that either sole into or are cut by the master detachment fault. After breakup of the hanging wall began, the master detachment fault excised part of the hanging wall and cut hanging-wall structures. The structural geometry of the Bannock detachment system strongly resembles that of detachments documented in metamorphic core complexes. Therefore, we interpret the Bannock detachment system as a proto-metamorphic core complex, akin to the Sevier Desert detachment fault.

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[1] Coauthored by Stephanie M. Carney and Susanne U. Janecke.

Two laterally extensive anticlines were produced by extension of the Bannock detachment system. The first, the Oxford Peak anticline, is an ESE-plunging anticline formed in a cohesive hanging wall of the master detachment fault during initial translation on the detachment fault system. Three-dimensional inclined shear may have produced the anticline. The second, the Oxford Ridge anticline, trends NNW, parallel to the strike of the Bannock detachment system, and likely formed due to isostatic rebound of the footwall of the master detachment fault. We interpret the Oxford Ridge anticline to be the proto-core of the Bannock detachment system.

The Bannock detachment system collapsed a structural culmination of the Sevier fold-and-thrust belt. The Cache-Pocatello culmination was a broad regional anticline that coincides in its N-S extent with the Bannock detachment system. Slip on the Bannock detachment system collapsed the Cache-Pocatello culmination in the same way that the Sevier Desert detachment collapsed the Sevier culmination (e.g., DeCelles et al., 1995).

Structures of the Bannock detachment system are overprinted by a second episode of extension accommodated by E- and NE-trending normal faults that may be related to subsidence along the Yellowstone hotspot track and a third episode of extension accommodated by high-angle, Basin-and-Range normal faults. This last episode of extension began no earlier than 4-5 Ma and continues today. The footwall of the Bannock detachment system was uplifted and exposed for the first time by the Basin and Range faults.

INTRODUCTION

Low-angle normal faults (detachment faults) are important yet enigmatic structures in extending regions. They accommodate large amounts of horizontal strain so they are often the principal structure of interest in extensional terranes (Allmendinger et al., 1983; Wernicke, 1995). Low-angle normal faults are regionally extensive features widely recognized in the Cordillera of the western United States (Longwell, 1945; Armstrong, 1972; Proffett, 1977; Allmendinger et al., 1983). However, despite many years of study there are persistent questions about the original dip of low-angle normal faults. Whether or not these faults are active and accommodate slip at dip angles less than 30° has been a controversial issue (Wernicke et al., 1985; Spencer, 1985; John, 1987; Jackson and White, 1989; Yin, 1989; Miller and John, 1999; Brady et al., 2000). Andersonian fault mechanics models do not predict slip on normal faults with angles of dip < 30° (Anderson, 1951), and few earthquakes have been identified on low-angle structures in extending regions (Jackson and White, 1989; Abers, 1991; Abers et al., 1997; Boncio et al., 2000). However, the geologic record in areas such as the Basin-and-Range province of the western United States provides conclusive evidence for past activity on low-angle normal faults (John, 1987; Lister and Davis, 1989; Scott and Lister, 1992; John and Foster; 1993; Miller and John, 1999; Brady et al., 2000).

Detachment-fault systems are defined as low-angle normal faults that are sub-regional to regional in extent and exhibit large amounts of horizontal displacement (typically >10 km) (Wernicke, 1992, 1995). Detachment systems are commonly associated with highly extended metamorphic core complexes (Crittenden et al., 1980), but are also documented in less extended terranes. In the model of Davis and Lister (1988), these extensional systems are high-angle (>30°) listric normal faults at the breakaway and flatten with depth into the mid-crust. With continued extension, the hanging wall of the system breaks up along planar, low-angle normal faults as well as planar and listric high-angle normal faults that are either cut by the master fault or sole into it (A in Fig. 3-1). As the crust in the hanging wall extends and overburden thickness decreases, mid-crustal rocks in the footwall, as well as the bounding master fault itself, dome upward to create a broad antiform. During the doming process, the master fault sometimes excises some of the hanging-wall strata (John, 1987; Davis and Lister, 1988). Break up of the hanging-wall creates small, localized basins that are characterized by gently dipping, locally derived, syntectonic strata (Fedo and Miller 1992; Friedman and Burbank, 1995). Low cut-off angles to the low-angle normal fault are diagnostic of an original low dip of the fault (e.g. Scott and Lister, 1992).

Although many models interpret planar or listric normal faults in the hanging wall as coeval with and soling into an underlying detachment fault, the low-angle master fault may be the youngest fault in the detachment system of a metamorphic core complex if the process of excision is important (excision is defined in Davis and Lister, 1988). During excision, the master low-angle fault steps up into its hanging wall and cuts both strata and structures in the hanging wall. Large amounts of slip are required after excision to exhume and erode the lens of faulted hanging wall between the initial and final positions of the detachment fault (Fig. 3-2).

The rotation of high-angle faults to lower angles during extension has been referred to as 'domino-style' extension (B in Fig. 3-1) (Proffett, 1977; Wernicke and Burchfiel, 1982; Gans and Miller, 1983). This style of extension is often localized above a detachment fault (e.g. Miller et al., 1983), but a basal detachment is not required (Proffett, 1977). In a domino-style extensional system without a basal detachment fault, high-angle, planar or listric, normal faults rotate and are cut by younger moderately to steeply dipping normal faults (Proffett, 1977). In a domino-style fault system, pre-rift and synrift sedimentary rocks have high cut-off angles to the faults (Proffett, 1977). Extension of the area is taken up by younger, high-angle normal faults as older hanging-wall faults rotate to angles less than 30° and become inactive as the area extends.

Another model to explain the origin of low-angle normal faults is the rolling-hinge model (C in Fig. 3-1) (Buck, 1988; Wernicke and Axen, 1988). In this model isostatic uplift of the footwall of a detachment surface occurs due to crustal unloading before and after slip on a series of moderate to steeply dipping, normal faults in the hanging wall (Buck, 1988; Wernicke and Axen, 1988). These hanging wall faults rotate to lower angles due to flexure and tilting caused by the rebounding crust in the footwall. As extension continues, the hinge of the footwall migrates in the direction of extension, and tilts fault blocks and the master fault to low angles as it migrates.

In this paper, we examine the geometry and evolution of an extensional system, the Bannock detachment system, located at the eastern edge of the Basin-and-Range province. We carefully examine the structural characteristics of this regional system of NNW-striking, WSW-dipping, low-angle normal faults in the Bannock Range of southeast Idaho. We will first describe the critical geometry and relationships along this fault system, then investigate the structure of the footwall and hanging wall, and finally interpret the origin and evolution of these structures in light of the controversy about low-angle normal faults. It is beyond the scope of this paper to solve the mechanical paradox of low-angle normal faults. Instead, we present the data and inferences that lead us to interpret an original low dip angle for the master fault of the Bannock detachment system. The along-strike equivalents of the major extensional system described here have been interpreted farther north to result from domino-style rotation of fault blocks, in the absence of a master detachment fault (Rodgers et al., in press). Our analysis therefore will focus on considering whether domino-style extension or a detachment model best explains the relationships in our study area.

INTRODUCTION TO THE STUDY AREA

Regional Tectonic Setting

The Bannock Range is within the Sevier fold-and-thrust belt of the North American Cordillera (Armstrong, 1972; Allmendinger, 1992), and lies near the northeastern margin of the Basin-and-Range extensional province in southeast Idaho (Fig. 3-3). The Sevier thrust-belt structures just east of the study area trend north, but curve to a northwest trend as they near the younger Eastern Snake River Plain. This change in trend occurs roughly where the Portneuf and Bear River ranges meet (Fig. 3-4). The Bannock Range is just east of the Malad ramp of the Paris thrust (Rodgers and Janecke, 1992; Platt, 1991).

At least two phases of extension followed Sevier orogenic contraction in the western United States (Wernicke, 1992; Stewart, 1998; Janecke et al., in press; Chapter 2 of this thesis). The first phase of extension was accommodated by low-angle normal-fault systems in both the hinterland and fold-and-thrust belt of the Sevier Orogeny. Extension in the hinterland is characterized by sequential development of latest Eocene to Miocene metamorphic core complexes (Miller et al., 1987; Miller and John, 1999; Miller et al., 1999; Mueller et al., 1999; Wells et al., 2000). This extension was localized in a belt west and southwest of the study area and has been well documented in the Raft River, Grouse Creek, and Albion Ranges of northern Utah and southern Idaho (Miller, 1991; Wells et al., 2000). Paleogene reactivation of thrust faults as normal faults has been documented in the Sevier fold-and-thrust belt in southwestern Wyoming (Royse et al., 1975; West, 1992,1993; Constenius, 1996). The Sevier Desert detachment, in central Utah, collapsed a culmination in the Sevier thrust belt (DeCelles et al., 1995). This detachment fault accommodated a large amount of extension, > 39 km, though no metamorphic core complex developed with the system (Coogan and DeCelles, 1996). The structural style of the proposed Bannock detachment system was similar to that of this first phase of extension.

The second phase of extension is characterized by moderately to steeply dipping, planar to listric normal faults that define the present basins and ranges in the province. The Bannock Range and adjacent Malad Range are both bounded on the east and west by north-striking, active normal faults. The west-bounding fault to the Malad Range is the Wasatch fault, which is an active normal fault that extends southward along the Wasatch front in northeastern Utah (Machette et al., 1992) (Fig. 3-4). Map patterns show that the normal faults in the Basin-and-Range province follow the gross geometry of the Sevier fold-and-thrust belt and change from dominantly north trends in the south to northwest trends farther north along a NE-trending line ~100 km southeast of the margin of the eastern Snake River Plain (Figs. 3-3, 3-4).

Several models have been proposed that indicate that the migration of the Yellowstone hotspot directly affected extension in the western U.S. (Anders et al., 1989; Rodgers et al., 1990; Pierce and Morgan, 1992; McQuarrie and Rodgers, 1998; Janecke et al., 2000). Initiation of hotspot activity in southwest Idaho and north-central Nevada occurred ~16 Ma (Pierce and Morgan, 1992) and was roughly coeval with the inception of Basin-and-Range extension in that area at about 17 Ma (Stewart, 1998). Studies by Anders et al. (1989), Rodgers et al. (1990) and Pierce and Morgan (1992) suggest that extension north and south of the hotspot track migrated north-eastward ahead of or coeval with the migrating hotspot. Because the edge of the Eastern Snake River Plain is approximately 120 km north of our study area (Fig. 3-3), our study area may have been influenced by the migrating hotspot. Morgan and Pierce (1999) showed that a 200-km-long zone of calderas in the central Snake River Plain was active in the Late Miocene about 10 Ma. They noted that the activity of the calderas was coincident with large-scale extension in the Raft River Valley, west of our study area (Fig. 3-3). Magmatic activity in this 200-km-long zone of calderas may also be coincident with a pulse of activity on the Bannock detachment system, which began during the late Miocene (~12 Ma) and continued into the Pliocene (Janecke and Evans, 1999; Janecke et al., in press; Chapter 2 of this thesis).

The younger Basin-and-Range normal faults in the study area fall into the Belt III zone of the projected seismic parabola around the Yellowstone hotspot (Pierce and Morgan, 1992) (Fig. 3-3). Belt III is characterized by waning fault activity.

Geologic Setting of the Study Area

The study area is in the southern part of the Bannock Range, and comprises the 7.5-minute Clifton quadrangle of Idaho (Figs. 3-4, 3-5). To the northeast and east of the Bannock Range are the southern Portneuf Range and the Bear River Range, respectively. South and west of the Bannock Range is the Malad Range (Fig. 3-4). The Clifton quadrangle is largely comprised of the Clifton horst, which is bounded to the east and west by north-trending, Basin-and -Range normal faults (Fig. 3-5). The Dayton-Oxford fault bounds the horst to the east, whereas the Deep Creek fault bounds the horst to the west. The Dayton-Oxford fault zone is a north-striking, east-dipping normal fault with at least 2 strands. It extends >10 km south and ~10 km north of the study area (Williams, 1962; Link, 1982a; Oaks, 2000) (Table 3-1). This fault zone bounds the west-tilted part of Cache Valley in and east of the study area (Janecke and Evans, 1999). The Deep Creek fault is a north-northwest-striking, west-dipping normal fault that bounds the west side of the Clifton horst (Janecke and Evans, 1999) (Table 3-1). This fault forms most of the east side of the Deep Creek half-graben, which contains mostly folded and faulted Tertiary and Quaternary rocks (Fig. 3-5) (Janecke and Evans, 1999; Janecke et al., in press; Chapter 2 of this thesis).

Strata in the Clifton horst consist of Neoproterozoic, early Paleozoic, Tertiary, and Plio-Pleistocene sedimentary, meta-sedimentary and igneous rocks. Structures exposed in the horst include numerous normal faults, several of which are low-angle, and small to large extensional folds.

Previous Work

Early work in the southern Bannock Range includes mapping and structural studies by Murdock (1961), Raymond (1971), Mayer (1979), Link (1982a, 1982b) and Oriel (unpublished). Several of these studies document younger-on-older normal faults (interpreted as younger-on-older thrusts by Raymond [1971] and Mayer [1979]) that, in some areas, place Cambrian rocks on Late Proterozoic rocks and that are also cut by more recent, high-angle, Basin-and-Range normal faults. Link (1982a, 1982b) documented low-angle normal faults in the Malad Summit and Oxford quadrangles north of our study area. He suggested that the low-angle normal fault system there resembled other low-angle fault systems (i.e., metamorphic core complexes) in the Basin-and-Range province. He also suggested that the low-angle normal fault system was late Mesozoic in age (Link, 1982). Recent work by Janecke and Evans (1999), Evans (in prep.), and Janecke et al. (in press; Chapter 2 of this thesis) documents that the low-angle normal faults are late Cenozoic because they place late Cenozoic rocks on Neoproterozoic rocks. These faults also place early Paleozoic rocks on Neoproterozoic rocks. Janecke and coworkers also recognize the regional extent of these older, low-angle normal faults, which they named the Bannock detachment system. The focus of their studies was mainly the Miocene-Pliocene Salt Lake Formation in the Deep Creek half-graben and Malad Range.

The Salt Lake Formation is a widespread, tuffaceous sedimentary deposit in the greater Cache Valley area (Fig. 3-4). It has been studied in detail by many workers (e.g., Danzl, 1985; Sacks and Platt, 1985; Brummer, 1991; Smith, 1997; Goessel, 1999; Goessel et al., 1999; Janecke and Evans, 1999; Oaks et al., 1999; Crane, 2000; Oaks, 2000; Pope and Link, 2000). The Miocene-Pliocene Salt Lake Formation is syntectonic to slip of the low-angle normal faults exposed in the study area (Janecke and Evans, 1999; Evans, in prep; Janecke et al., in press; Chapter 2 of this thesis). The Tertiary Salt Lake Formation filled a basin(s) in the hanging wall of the Bannock detachment system. The Valley fault of Sacks and Platt (1985), east of the study area in the Portneuf Range, is the breakaway for the detachment system (Janecke and Evans, 1999). Janecke and Evans (1999) and Janecke et al. (in press) presented a three-stage evolution of the northern Cache Valley area. They propose that the area experienced WSW-directed extension on the NNW-trending Bannock detachment system (stage 1) during the late Miocene-Pliocene, and was later overprinted by N-S striking Basin-and-Range normal faults (stage 3). Cross faults (stage 2) with NE, NW, and E-W strikes formed mostly between stage 1 and 3 (Janecke and Evans, 1999).

Our detailed mapping and structural analysis of the Clifton quadrangle is the basis for a greatly expanded understanding of the structure and evolution of the Bannock detachment system. In this paper we show that the Bannock detachment system: 1) accommodated moderate to large magnitudes of, WSW-directed extension, 2) was active at a low angle, 3) probably formed at a low angle, 4) was a regionally extensive system that produced large NNW- and ESE-trending anticlines, and 5) resembles a proto-metamorphic core complex, as first noted by Link (1982a).

METHODS

Detailed geologic field mapping with the aid of aerial photographs at a scale of 1:16,000 and age determinations based on geochemical correlations of tephras from the Salt Lake Formation were used to construct the tectonic and structural evolution of the southern Bannock Range. The results of the tephra correlations and stratigraphic and sedimentological analyses are reported in Janecke et al. (in press) and Chapter 2 of this thesis. A geologic map of the Clifton quadrangle at a scale of 1:24,000 as well as six geologic cross sections at the same scale were produced (Plate 1). A regional geologic cross section was also constructed across the entire extensional system. Structural data from published maps and theses were compiled on a 1:100,000-scale map that extends from the Samaria Mountains to the Portneuf Range of southeastern Idaho. This was reduced to produce Fig. 3-4 (see the figure caption for sources of the data). We revised a sub-Tertiary subcrop map of southeast Idaho from Rodgers and Janecke, (1992). We expanded the subcrop map into north-central Utah and incorporated new, detailed geologic mapping in the area of study and reconnaissance work in the greater Cache Valley area. The subcrop map encompasses an area from Pocatello, Idaho, to the south-central area of Cache Valley, Utah. Bedding attitudes were used to define domains of consistent strike and dip in Tertiary and pre-Tertiary rocks in the Clifton quadrangle. The resultant dip-domain map helped to identify and characterize several extensional folds and highlight the structural differences between the hanging wall and footwall of the Bannock detachment system. Stereonets were made to analyze extensional folds. Tables 3-1 and 3-2 summarize the geometry and origin of the normal faults and folds in the area and Table 3-3 summarizes the extensional events of the area.

STRATIGRAPHY

Strata in the study area from oldest to youngest consist of: Neoproterozoic Pocatello Formation, Neoproterozoic to Cambrian Brigham Group rocks, Cambrian to Silurian miogeoclinal carbonates and shales, Paleocene-Eocene Wasatch Formation, Miocene-Pliocene Salt Lake Formation, Late Miocene(?) mafic intrusions, and Pliocene-Pleistocene(?) piedmont gravel deposits (Fig. 3-6). The Pocatello Formation consists of greenschist facies metabasalts, volcanic breccia, diamictite, sandstone, siltstone, and conglomerate (Link, 1982a, 1982b; Link et al., 1993). These rocks are best exposed on Oxford Ridge and along the east side of the Clifton horst (Fig. 3-5). Stretched pebbles occur in diamictites of the Pocatello Formation and are thought to have formed coeval with metamorphism (Link, 1982a, 1982b). Brigham Group rocks consist largely of siltstone, coarse pebbly conglomerate with argillite interbeds, vitreous sandstone, pebble conglomerate, and micaceous mudstone with interbedded quartzite (Link et al., 1987). Strata of the Brigham Group are exposed in the Clifton horst and east of the horst just south of the town of Clifton (Fig. 3-5). Lower Paleozoic miogeoclinal strata consist mostly of limestones and dolostones, with some sandstones, quartzites, and shales. These rocks are exposed in the study area along Oxford Ridge and in the Clifton horst. The Tertiary Wasatch Formation, which is a synorogenic conglomeratic deposit of the Sevier fold-and-thrust belt (Oaks and Runnells, 1992), crops out in the study area in six isolated fault blocks on the northwestern margin of the Clifton horst west of Oxford Ridge (Fig. 3-5) where it is deposited on Ordovician and Cambrian strata.

The Tertiary Salt Lake Formation is widespread in the greater Cache Valley area and crops out extensively in the study area (Janecke et al., in press; Chapter 2 of this thesis). Four distinct members of the Salt Lake Formation are present in the Deep Creek half-graben west of the study area (Janecke and Evans, 1999; Evans, in prep; Janecke et al., in press; Chapter 2 of this thesis). We use the nomenclature of Janecke and Evans (1999) for members of the Salt Lake Formation because of the close proximity of their study area to ours and because their unit descriptions closely match ours. The basal Skyline Member consists mostly of poorly sorted pebble-to-cobble conglomerate with a tuffaceous, calcareous, and sandy groundmass and interbeds of lacustrine limestones and variable amounts of tephra. This unit is interpreted as an alluvial-fan deposit (Janecke and Evans, 1999). The Skyline Member was probably never deposited in the study area because it was restricted to an early-formed half graben in the west half of the younger Deep Creek half graben (Janecke et al., in press; Chapter 2 of this thesis). The Cache Valley Member consists of slightly reworked zeolitized and tuffaceous sedimentary rocks, including mudstones, siltstones, some limestones and silicified limestones, shales, sandstones, and rare pebble conglomerates, as well as some primary ash-fall tuffs and tephras. It is interpreted as a saline/alkaline lacustrine deposit (Janecke and Evans, 1999; Janecke et al., in press; Chapter 2 of this thesis). The Third Creek Member overlies and intertongues with the Cache Valley Member. It is composed mostly of medium- to coarse-grained tuffaceous sandstones, clast- and groundmass-supported conglomerates, and poorly consolidated white to silvery-gray tephra. It is interpreted to consist of nearshore-, fluvial-, and deltaic deposits (Janecke and Evans, 1999; Janecke et al., in press; Chapter 2 of this thesis). Overlying the Third Creek Member is the New Canyon Member, which is poorly consolidated in the study area. However, exposures northwest of the study area show that it is a parallel-bedded pebble-to-cobble conglomerate (Janecke and Evans, 1999). This member is interpreted to be a fluvial deposit (Janecke and Evans, 1999; Janecke et al., in press; Chapter 2 of this thesis).

Stratigraphic analyses indicate that the Salt Lake Formation is correlative across the younger Clifton horst (Janecke et al., in press; Chapter 2 of this thesis). The Cache Valley Member is interpreted to have been deposited in a broad, coherent basin because basin-margin facies are lacking (Janecke et al., in press; Chapter 2 of this thesis). The Third Creek and New Canyon Members, which contain marginal facies, are interpreted as deposits of more localized half graben resulting from breakup of the hanging wall of the Bannock detachment system

(Janecke et al., in press; Chapter 2 of this thesis). Altogether more than 2 km of Salt Lake Formation are preserved in the study area (Janecke et al., in press; Chapter 2 of this thesis).

A mafic, igneous sill, is exposed on Oxford Ridge in the north area of the study area. It has been metamorphosed to amphibolite facies, is fine to medium crystalline, and is largely composed of hornblende and plagioclase with lesser amounts of epidote, chlorite, sericite, leucoxene, hematite, and clay (Raymond, 1971). Crystals are 2 mm to 1 cm across.

Pliocene-Pleistocene(?) piedmont gravel deposits are preserved as local, dissected erosional remnants within the Clifton horst, and as widespread piedmont deposits in two areas west of the horst (Fig. 3-5). The lithology, clast size, composition, and degree of cementation of these conglomeratic deposits depend strongly on the lithology of the adjacent and underlying bedrock. These deposits partially bury the proto Clifton horst during an interval of tectonic quiescence in extensional episode 3 (Janecke et al., in press; Chapter 2 of this thesis) (Table 3-3).

STRUCTURE

The structure of the southern Bannock Range is complex. Crosscutting relationships in the study area show that three, distinct sets of normal faults and extensional folds record different stages of extension (Table 3-3). The oldest normal-fault set is related to the sequential development of the Bannock detachment system. We used crosscutting relationships to distinguish four subphases of set 1. The main subset of fault set 1 (1b) is made up of low-angle normal faults of the Bannock detachment system. There may be older localized normal faults (set 1a) that controlled the deposition of the basal conglomerate of the Tertiary Salt Lake Formation in the Cache Valley region (Sacks and Platt, 1985; Janecke and Evans, 1999; Goessel et al., 1999). A major ESE-plunging anticline probably formed in the hanging wall of the Clifton fault during initial translation on the master detachment fault. The folded hanging wall of the detachment fault broke up internally along WSW-dipping normal faults (set 1c) that soled into the master detachment fault. Excision of the hanging wall occurred late in the evolution of the detachment system when the Clifton fault cut up into it (set 1d). Low-angle normal faults of fault set 1b, 1c and 1d are exposed in the Clifton horst and east of the horst. Fault set 1 accommodated the bulk of the extension in the area.

The second fault set is made up of cross-faults, which cross-cut the older low-angle normal faults. These high-angle normal faults strike E and NE, and dip N, NW, and S. These faults are exposed in the southern part of the Clifton horst (Fig. 3-5). These faults accommodated small amounts of extension.

The third and youngest fault set is made up of high-angle, N-trending, E- and W-dipping, Basin-and-Range normal faults (Table 3-3). These faults bound the Clifton horst on the east and west and, locally, some break up the horst (Fig. 3-5). Faults of this set are very common in the Deep Creek half-graben, west of the study area. Modest amounts of extension accumulated on faults of set 3.

Pre-extension Structures

Extension in the study area occurred after the end of contraction associated with the Sevier Orogeny. By mapping the unconformity between the lower Paleozoic strata and the Tertiary Wasatch and Salt Lake Formations in the study area, as well as regionally, we constructed a subcrop map that reflects the pre-extensional structural relief in southeast Idaho and north-central Utah, following the methods of Armstrong (1968) and Rodgers and Janecke (1992) (Figs. 3-7). Structures on the subcrop map formed during thrusting and folding of the Sevier Orogeny. Our subcrop map expands and updates the work of Rodgers and Janecke (1992) based on our new mapping around in northern Cache Valley, recent compilations (Link and Stanford, 1999) and mapping in other areas of the region (Platt, 1977; Oriel and Platt, 1980; Link, 1982a; Danzl, 1985; Dover, 1995; Kellogg et al., 1999; Link and Stanford, 1999; Crane, 2000; Pope et al., 2001). Our study area is in the hanging wall of the Paris thrust, above a footwall and hanging wall flat, and east of the buried Malad ramp (a footwall ramp) of this thrust (Rodgers and

Janecke, 1992). Their study shows evidence for a broad, NNW-trending anticline that developed above an upper-flat of the Paris thrust. Our subcrop map also shows a broad NNW-trending uplifted area, but unlike Rodgers and Janecke (1992), the oldest rocks exposed in the core of the contractional anticline are of the Cambrian Blacksmith Formation (Fig. 3-7) and not rocks as old as the Neoproterozoic Pocatello Formation (e.g., Link, 1982a, 1982b). Our new mapping reveals that previously mapped depositional contacts between the Tertiary Salt Lake Formation and the underlying Neoproterozoic-Cambrian Brigham Group and Neoproterozoic Pocatello Formation are fault contacts (Figs. 3-4, 3-5).

Several key locations define the extent of the NNW-trending anticline between the Malad ramp to the west and the Logan Peak syncline to the east. West of the study area, in the Samaria Mountains west of the Malad ramp of the Paris thrust, rocks below the unconformity young westward and are as young as the Permian (Platt, 1977). Just west of the study area, in the Malad Range, Silurian and Ordovician rocks are overlain by the Salt Lake Formation (Evans, in prep). In the northwest part of the study area, the Ordovician Garden City Formation is unconformably overlain by the Tertiary Salt Lake Formation (Fig. 3-5). East of the study area, in the Swan Lake block of the Portneuf Range, rocks as old as the Cambrian Blacksmith Formation are in angular unconformity with Tertiary rocks (Fig. 3-4) (Sacks and Platt, 1985; Danzl, 1985). Rocks under the unconformity young eastward from Cambrian to Silurian in the Portneuf Range and are as young as Pennsylvanian to Permian in the core of the Logan Peak syncline of the Bear River Range (Dover, 1995). The anticline presently has a north-south extent of approximately 135 km and an east-west extent of 30-40 km (Fig. 3-7). It extends as far north as the Snake River Plain and as far south as central Cache Valley, approximately the latitude of Logan, Utah (Fig. 3-3). The most uplifted part of the anticline is in the southern Portneuf Range. We refer to this anticline as the Cache-Pocatello culmination after Janecke et al. (in press).

BANNOCK DETACHMENT SYSTEM
(FAULT SET 1)

The Bannock detachment system is made up of multiple low-angle-fault strands with the Neoproterozoic Pocatello Formation in the footwall. The attitudes of these faults vary across the area, probably due to original corrugations and/or later warping, but overall the detachment system has NNW strikes and gentle, WSW-dips (Fig. 3-5; Table 3-1). We interpret the Clifton fault as the master low-angle normal fault because it is the most continuous structure in the study area and has the greatest stratigraphic omission. In several places, the Clifton fault juxtaposes Tertiary Salt Lake Formation with Neoproterozoic Pocatello Formation, and there omits >2 km of strata (Fig. 3-5). There are several low-angle normal faults in the footwall of the Clifton fault that are best exposed on the east face of Oxford Ridge (Fig. 3-5) and there are several low-angle and high-angle normal faults in the hanging wall of the Clifton fault. Except for younger Basin-and-Range faults, none of these faults cuts the Clifton fault but several merge and sole into it. The Clifton fault and all low-angle faults in its footwall comprise the Bannock detachment system and represent fault set 1b (Table 3-3). All faults that sole into or are cut by the Clifton fault are considered hanging-wall structures to the Bannock detachment system and represent fault set 1c (Table 3-3).

Clifton Low-angle Normal Fault

The Clifton fault strikes NW along most of the Clifton horst and is exposed for ~9 km south and ~14 km north of the study area (Link, 1982a, 1982b; this study). It persists at least 39 km N-S. The fault has a low-angle geometry through its entire extent, has an average dip of 15° to the WSW, and exhibits gentle waviness along strike (Fig. 3-8). Foliated and massive rocks of the Neoproterozoic Pocatello Formation comprise the footwall of the Clifton fault, whereas hanging-wall rocks vary along strike from strata of the Neoproterozoic to Cambrian Brigham Group, Cambrian to Ordovician Formations and the Tertiary Salt Lake Formation. Slip direction of the hanging wall is probably to the WSW based on foliation dips in the Pocatello Formation that are perpendicular to the strike of the Clifton fault (Link, 1982b; this study) and based on the northwest strike of normal faults that sole into the Clifton fault.

Slip on the Clifton fault was calculated to be greater than 15 km using cross section B-B' (Fig. 3-9a). There the Clifton fault dips ~15° WSW and juxtaposes Ordovician rocks with Neoproterozoic Pocatello Formation. Cross section B-B' was restored to remove younger Basin-and-Range faulting (Fig. 3-9b). The minimum amount of slip on the Clifton fault based on overlap of dissimilar rocks is ~4 km. By expanding cross section B-B' eastward to project a footwall cutoff to the Clifton fault at a minimum distance, we calculated ~6 km more slip. We calculated that at least another ~5 km of slip occurred during excision of the hanging wall by the Clifton fault. The total minimum slip of the Clifton fault is ~15 km.

On Oxford Ridge in the north of the study area, the hanging wall of the Clifton fault is composed mostly of a very thin fault blocks of Cambrian through Ordovician strata (Fig. 3-9a, b). Several closely spaced, NW-striking normal faults cut these lower Paleozoic rocks, but do not cut the Clifton fault (Figs. 3-5, 3-9a, b). It appears that these faults have been cut by the Clifton fault because they have fairly steep dips and are difficult to balance in reconstructions (Fig. 3-9a, b).

However, the Clifton fault is offset on the east side of the Clifton horst by strands of the Dayton-Oxford fault, a high-angle Basin-and-Range fault, and is also cut by two E- and NE-striking cross faults in the southern part of the Clifton horst (Fig. 3-5). The Clifton fault is interpreted to have been offset in the subsurface by the Deep Creek fault (Figs. 3-9, 3-10, 3-11, 3-12), but this crosscutting relationship is not exposed.

The Clifton fault and its footwall are folded into a broad, open anticline, the Oxford Ridge anticline, with an axis trending roughly NNW (Table 3-2). Evidence for this fold is best seen on Oxford Ridge. There, to the immediate west of Oxford Ridge, the Clifton fault dips about 15° west, but changes to a very gentle east dip on the east edge of the ridge (Fig. 3-9a, b). Other low-angle normal faults of the Bannock detachment system in the footwall of the Clifton fault are also folded by the NNW-trending anticline (Fig. 3-13). These faults and the Oxford Ridge anticline are described in more detail below. Foliations in the Pocatello Formation define the anticline (Fig. 3-14).

Footwall of the Bannock Detachment System

The Neoproterozoic Pocatello Formation makes up the footwall of the Bannock detachment system in the area of study. The volcanic and sedimentary rocks of the Pocatello Formation have undergone low-grade metamorphism (Link, 1982a, 1982b; Link et al., 1993) and were not well bedded prior to deformation. Therefore, detecting the attitude of bedding is difficult. Sedimentary beds in the Pocatello Formation dip east at approximately 35° along the steep eastern face of the Clifton horst (Fig. 3-5) (Link, 1982a). No detectable bedding was found to the west of Oxford Ridge in the west limb of the Oxford Ridge anticline, but the rocks may dip west based on the map pattern of the Scout Mountain Member versus the underlying Bannock Volcanic Member of the Pocatello Formation along Second Creek (Fig. 3-5 point C). North of the study area in the Oxford quadrangle (Fig. 3-4), the Pocatello Formation is locally mylonitic beneath the Clifton fault (Link, 1982a, 1982b). Mylonitic fabrics also coincide with some low-angle faults beneath the Clifton fault there. No mylonitic fabrics are observed in the study area.

Flattened and stretched pebbles of diamictites in the Pocatello Formation are documented in numerous locations along the Clifton horst. Analysis of the direction of the long axis of 40 pebbles indicates a dominant strain direction of ~288° (Fig. 3-15). The strain direction recorded by the stretched pebbles is ~40° clockwise from the WSW slip direction inferred along the WSW-dipping Clifton fault. These pebbles may have been stretched during Sevier-aged shortening and/or during faulting on the Bannock detachment system. Further study is needed to assess their age and kinematic significance.

Foliation in the Pocatello Formation is observed in many locations in the Clifton horst. However, some outcrops are massive and unfoliated. The foliation changes dip direction from WSW to ESE eastward across the Oxford Ridge anticline (Fig. 3-5). At Oxford Ridge, the axial trace of the anticline is ~30° counterclockwise from the steep range-front fault, the Dayton-Oxford fault, and the anticline is cut by this fault farther south near Clifton Basin (Fig. 3-5). In the south part of the study area, the dip of foliation is dominantly to the WSW and SW and defines the west limb of the anticline. In the Oxford quadrangle to the north, the exposures of the Pocatello Formation are restricted to the east limb of the Oxford Ridge anticline. Foliation and bedding in the Pocatello Formation there dips to the east and northeast (Link, 1982a, 1982b). The east limb of the anticline is faulted down into Cache Valley, but it is exposed east of the study area in a horst block at Little Mountain (Fig. 3-4). Foliation in the Pocatello Formation there has an E to SE dip direction (Link and LeFebre, 1983).

The Oxford Ridge anticline is best exposed on Oxford Ridge in the north half of the study area. It can be traced for more than 16 km in the study area and more than 40 km regionally (Figs. 3-4, 3-5). The anticline trends 164°, sub-parallel to the strike of exposures of low-angle normal faults of the detachment system, and plunges 0° based on the foliation data (Fig. 3-14 and Table 3-2). The east limb of the anticline dips 27° E and the west limb dips 21° W (Fig. 3-14). The Clifton detachment fault is subparallel to foliation in the Pocatello Formation and is folded about as much as the foliation (Figs. 3-5, 9a, b).

At least 3 low-angle faults are exposed on the steep east face of Oxford Ridge below the Clifton fault and one is exposed on the west side of Oxford Ridge (Figs. 3-5, 3-16). These faults are restricted to the Pocatello Formation and cut only the members of the Pocatello Formation and an undated, younger, mafic intrusion that we interpret as Neogene (see below). These low-angle faults are sub-parallel to each other and to the Clifton fault, but are structurally lower than the Clifton fault. They appear to dip gently east on the east-facing side of Oxford Ridge (Fig. 3-

13) and gently west where one of them is exposed at Second Creek west of Oxford Ridge (Fig. 3-5, point C). The low-angle fault exposed at Second Creek is definitely a normal fault. The fault dips west and juxtaposes the younger Scout Mountain Member of the Pocatello Formation in the hanging wall on the older Bannock Volcanic Member in the footwall. Therefore, we infer that all of these low-angle faults confined to the Pocatello Formation are normal faults, not thrust faults. We interpret them as bounding large thin lenses of rock in the subsurface (Fig. 3-9a, b). Like the foliation in the Pocatello Formation and the Clifton detachment fault, these low-angle normal faults are gently folded by the Oxford Ridge anticline.

A mafic sill, and smaller satellite bodies, intruded the Pocatello Formation in the footwall of the Clifton fault (Raymond, 1971; this study) (Fig. 3-5). This mafic sill crops out on Oxford Ridge and, like the Pocatello Formation, has been mildly deformed and metamorphosed up to amphibolite facies (Raymond, 1971). Exposures of the sill extend about 1.4 km from west to east and about 4 km from north to south (Fig. 3-5).

The mafic sill is in low-angle fault contact with the overlying Scout Mountain Member and underlying Bannock Volcanic Member of the Pocatello Formation. The sill is approximately 120 m thick (Fig. 3-17). It is foliated and sheared in the deepest outcrop exposure on Oxford Ridge, but is massive and faulted in shallower exposures. Mapping indicates that the mafic sill intruded along a low-angle normal fault and so that is underlies the Scout Mountain Member and overlies the Bannock Volcanic Member. Foliation in the sill dips gently southwest, parallel to sub-parallel to the low-angle faults and foliation in the host rock, the Pocatello Formation (Figs. 3-9a, b, 3-17). Small, unfoliated dikes from the mafic sill intruded into the overlying Scout Mountain Member. The upper-most part of the sill contains brittle fault rocks that are intruded by less deformed felsic restite of the mafic sill.

The mafic sill intruded along a low-angle normal fault within the Pocatello Formation during extension and exhumation of the footwall of the detachment system. It is probably

Neogene in age because it intrudes a low-angle normal fault of the Bannock detachment system and its pre-metamorphic lithology is similar to that of unmetamorphosed diabases that cut the Tertiary Salt Lake Formation and intruded along normal faults in NE Cache Valley (Winter, 1989) (Fig. 3-4). Similar mafic intrusions were documented south of the study area in the Clarkston and Portage quadrangles of northern Utah and southern Idaho and in the southwest corner of the Weston Canyon quadrangle of southern Idaho (Prammani, 1957; Biek et al., 2001). One intrusion there that invaded the Salt Lake Formation yielded a K-Ar age of 8.0 ± 0.5 Ma (date of D. Fiesinger, cited in Biek et al., 2001). Altogether, these relationships show cooling coincident with deformation of the mafic sill and are consistent with emplacement during extensional exhumation of the footwall of the Clifton fault in the late Cenozoic. The amphibolite facies metamorphism described by Raymond (1971) must also be late Cenozoic if the mafic sill is Miocene, as we infer.

Hanging Wall of the Clifton Fault

Oxford Peak Anticline. The oldest extensional structure in the hanging wall of the Bannock detachment system is an ESE-plunging anticline, the Oxford Peak anticline, defined by a ~96° change in strike of bedding in the Brigham Group and lower Paleozoic rocks in the hanging wall of the Clifton fault. The dip of these rocks in the southern area of the Clifton quadrangle, near and southeast of Old Baldy Peak, ranges from 20° to 45°, but is consistently to the SE (Figs. 3-5, 3-18). In the central part of the Clifton quadrangle, the Brigham Group rocks in the footwall of the Clifton Cemetery fault and the footwall of the Clifton Basin fault, as well as rocks near Buck Peak, also dominantly dip to the SE (Figs. 3-5, 3-18). This dominant SE dip of Brigham Group and lower Paleozoic strata continues to the northern edge of the map area and as far south as the north side of Dry Canyon, ~6 km south of the Clifton quadrangle. Approximately 2.3 km north of the map area, at Oxford Peak, the dip direction of Brigham Group rocks in the hanging wall of the Clifton fault changes from SE to dominantly E (Link, 1982a) (Fig. 3-4). From the Oxford Peak area to about 7.5 km north, the dip direction, again changes, from E to NE (Link, 1982a) (Fig. 3-4). The broad hinge of this plunging anticline is at Oxford Peak (Fig. 3-4).

In the Bannock Range, the dip domains that define this anticline are only present in the pre-Tertiary strata in the hanging wall of the Clifton fault. In the footwall of the fault, foliated strata dip either WSW or ENE on either limb of the Oxford Ridge anticline and commonly strike at a high angle to pre-Tertiary rocks in the hanging wall of the Clifton fault. Few exposures of the Tertiary Salt Lake Formation exhibit SE dips in the study area, but such dips are present along Mink Creek southeast of the southeastward projection of the fold (Janecke, unpublished data).

Northeast- and southeast-dipping limbs of the Oxford Peak anticline have a north-south extent of about 20 km and define a fold axis that trends 106° and plunges 24° ESE (Fig. 3-14, Table 3-2). The Oxford Peak anticline is buried beneath Cache Valley. However, it is noteworthy that the axial trace projects ESE toward a point near the intersection of the N-striking East Cache fault in the Bear River Range and the NW-striking Valley fault in the Portneuf Range (Fig. 3-4).

It is possible that the anticline is present to the east of our study area in the hanging wall of the Valley fault in the Portneuf Range (Fig. 3-4). There, exposed Brigham Group and lower Paleozoic strata have a dominant NE dip direction north of Mink Creek (Platt and Oriel, 1967; Sacks and Platt, 1985; Danzl, 1985) (Fig. 3-4). This dip domain is just north of the projected axis of the anticline. These strata may be part of the NE-dipping limb of the Oxford Peak anticline. Also, changes in the strike of Paleozoic rocks at the southeast end of the Swan Lake fault block (Fig. 3-4, data of Danzl, 1985) define a ESE-plunging anticline that is probably the ESE end of the Oxford Peak anticline. South of the projected axis of the Oxford Peak anticline, there are ESE-dipping Paleozoic rocks which may be part of the southern limb of the Oxford Peak anticline (Fig. 3-4).

It is uncertain whether this anticline continues west of our study area into the Malad Range (Fig. 3-4). That area is complicated by numerous normal faults and folds related to young

Basin-and-Range extension (Janecke et al., 1999; Evans, in prep.). Exposure of Brigham Group rocks is localized (Janecke et al., 1999; Evans, in prep.) and few consistent dip domains exist in sub-horizontal lower Paleozoic strata.

It is unlikely that the Oxford Peak anticline is a Sevier-aged structure. The axis of the Oxford Peak anticline trends and plunges ESE, unlike large N- and NW-trending folds in Paleozoic and older rocks near the study area that formed during the Sevier Orogeny. We distinguish between this regional ESE-trending anticline and NW- to NNE-trending structures of the Sevier fold-and-thrust belt and infer that the Oxford Peak anticline formed during Cenozoic extension after deposition of the Cache Valley Member of the Salt Lake Formation but before deposition of the Third Creek Member of the Salt Lake Formation. The pre-Tertiary subcrop map shows little evidence for the Oxford Peak anticline prior to deposition of the Tertiary Wasatch Formation and the Cache Valley Member of the Salt Lake Formation (Fig. 3-7).

Crosscutting relationships between lower Paleozoic and Brigham Group strata of the Oxford Peak anticline and low-angle normal faults in the hanging wall of the Clifton fault indicate that folding occurred after detachment faulting began, but before breakup of the hanging wall of the Bannock detachment system. In the study area, the SE-dipping limb of the Oxford Peak anticline lies above the Clifton fault, but comprises the footwall to several low-angle, hanging wall faults to the master (Clifton) detachment fault. Crosscutting relationships in the Clifton Cemetery fault block east of the Clifton horst show that the SE-dipping footwall rocks of the Brigham Group were tilted before faulting of the SW-dipping Clifton Cemetery fault began (Figs. 3-5, 3-10a, b).

Because this ESE-plunging anticline is confined to the hanging wall of the Clifton fault, and the subcrop map suggests that it postdates the oldest Tertiary rocks, we infer that it formed early during detachment faulting, probably during the translation phase on the detachment system (Table 3-3, episode 1b). We believe that the Oxford Peak anticline likely formed above a concave-up, lateral bend in the Bannock detachment system during WSW-directed extension of the master fault of the Bannock detachment system (see below).

Imbricates in the Hanging Wall of the Clifton Fault and Associated Extensional Folds. The hanging wall of the Clifton fault is also defined by a series of low-angle normal faults and moderately to steeply dipping, listric and planar normal faults that either are cut by the master fault or sole into it. These faults have a range of displacements and include both synthetic and antithetic NW- to NNW-striking normal faults. Synthetic faults are more numerous than antithetic faults (Fig. 3-5). Several extensional folds are confined to the hanging wall of the Clifton fault.

The hanging-wall faults in some areas are closely spaced, synthetic or antithetic, high-angle normal faults that strike parallel to or are sub-parallel to the master fault. An example of these hanging-wall features is on the west side of Oxford Ridge in the north part of the study area (Fig. 3-5). There the hanging wall of the Clifton fault is cut by numerous high-angle listric and planar normal faults spaced 200 m to 400 m apart (Figs. 3-5, 3-9a, b). They cut Cambrian and Ordovician strata as well as the Tertiary Wasatch and Salt Lake Formations. Offset is between 400 m and at least 700 m. The more prominent faults in this area are synthetic to and strike sub-parallel to the Clifton fault. However, none of these high-angle normal faults cut the Clifton fault.

Elsewhere larger normal faults, with much greater spacings, extend the hanging wall of the Clifton fault (Fig. 3-5). These fault blocks are spaced between 800 m and 2,800 m apart (Fig. 3-5). The Pocket Basin fault is a low-angle normal fault synthetic to the Clifton fault. It is exposed in the Clifton horst and merges with the Clifton fault just west of Oxford Ridge (Fig. 3-5, point A). It has a NW strike, sub-parallel to the Clifton fault, and has an average WSW dip of 18° (Table 3-1), though dips vary along strike because it exhibits a scoop-like geometry that is probably due to younger folding. The Pocket Basin fault defines the NW-trending Pocket Basin syncline. The Pocket Basin fault is inferred to sole into the Clifton fault at depth because it merges with the Clifton fault northward in the geologic map (Fig. 3-12).

The Pocket Basin fault places the Tertiary Salt Lake Formation in the hanging wall on Brigham Group rocks in the footwall (Figs. 3-5, 3-12). Minimum offset on this fault is >2.5 km based on the overlap of dissimilar rocks in the hanging wall and footwall (Fig. 3-12). The hanging-wall rocks are subparallel to the fault, whereas footwall rocks dip to the SE. The overall geometry of the Pocket Basin fault is that of a hanging-wall flat above a lateral footwall ramp. The cut-off angle between the hanging-wall rocks and the fault is very low, less than 5°, evidence of an original low dip of the Pocket Basin fault. As indicated earlier, the SE dip of the footwall strata predates the WSW-dipping Pocket Basin fault, and developed on the SE-dipping limb of the Oxford Peak anticline.

The Pocket Basin syncline is apparent in the Cache Valley Member of the Salt Lake Formation in its hanging wall and in the footwall of the younger Deep Creek fault (Figs. 3-5, 3-12). The axial trace of this syncline trends 313°, roughly parallel to the Pocket Basin fault but also sub-parallel to the younger Deep Creek fault, and it plunges 16° NW (Fig. 3-14, Table 3-2). Folded Brigham Group rocks in the footwall of the Pocket Basin fault indicate that the syncline extends at least 1.2 km southeast into the footwall of the Pocket Basin fault (Fig. 3-5). The northeast limb of the syncline is defined by SW-dipping strata with similar dips as strata in the west limb of the Clifton Basin anticline. It is possible that the Pocket Basin syncline was once part of the west limb of the Clifton Basin anticline and was folded into a syncline due to uplift of the footwall of the younger Deep Creek fault.

The Clifton Cemetery fault is exposed east of the Clifton horst two km south of the town of Clifton (Fig. 3-5). The Clifton Cemetery fault is cut by the younger East Dayton-Oxford fault (Fig. 3-5). It is exposed in the hanging wall of the Dayton-Oxford fault and in the footwall of the East Dayton-Oxford fault. It is a low-angle normal fault synthetic to and structurally higher than the Clifton fault. It strikes NW, between 11° and 29° counterclockwise to the Clifton fault, and has an average dip of 21° SW (Table 3-1). The Clifton Cemetery fault juxtaposes SE-dipping

Brigham Group rocks in the footwall and NE-dipping Tertiary Salt Lake Formation in the hanging wall. The hanging-wall rocks strike nearly parallel to the fault and dip gently into it. Minimum dip slip on this fault is more than 1.5 km (Fig, 3-10a, b). A NW-dipping normal fault of fault set 2 offsets the Clifton Cemetery fault along Clifton Creek (Fig. 3-5).

Several other hanging-wall faults are exposed near Weston Peak in the southern part of the Clifton horst (Fig. 3-5). Again, none of these normal faults cuts the Clifton fault. A major, listric normal fault, antithetic to the Clifton fault, was mapped near Old Baldy Peak (Fig. 3-5). This fault strikes NNE, dips to the ESE about 22°, and juxtaposes Cambrian strata in the hanging wall on Brigham Group rocks in the footwall (Fig. 3-10a, b). This fault appears to cut the southern continuation of the Pocket Basin fault and several smaller low-angle faults in the Old Baldy Peak area (Fig 3-5). However, this fault does not appear to cut the Clifton fault and likely soles into it (Fig. 3-10a, b). The hanging wall of this normal fault is cut by numerous, closely spaced normal faults that have a variety of strikes.

Another sizable, listric normal fault was mapped in Clifton Basin. It is here referred to as the Clifton Basin fault (Fig. 3-5). This curvi-planar normal fault is antithetic to the Clifton fault, trends roughly NNE, and dips to the ESE (Fig. 3-11a, b). It places SW- and NW-dipping, Tertiary Salt Lake Formation in the hanging wall on SE-dipping, Brigham Group rocks in the footwall (Fig. 3-11a, b). The hanging-wall strata next to the fault strike slightly oblique to the fault, but dip into it. The hanging-wall strata were deposited before slip on this fault. Map patterns show that the Clifton basin fault does not cut the Clifton fault, but soles into it or is cut by it (Fig. 3-5).

The hanging wall of the Clifton Basin fault has been folded into a SW-plunging anticline. The Clifton Basin anticline is best defined at Clifton Basin, but likely extends east of the Clifton horst. The anticline in Clifton Basin folds strata of the Salt Lake Formation. It is in the hanging wall of the Clifton fault, and soles into or is truncated by the Clifton fault (Fig. 3-11a, b). The axial trace of the anticline trends 217° and plunges 15° SW (Fig. 3-14, Table 3-2). The west limb of the anticline dips into the ESE-dipping Clifton Basin fault. Beds of the eastern limb in Clifton Basin dip as steeply as 64° SE next to the SW-dipping Clifton fault (Fig. 3-11a, b).

The southeastern limb of the Clifton Basin anticline is faulted down to the east by the Dayton-Oxford fault and may be represented by ENE-dipping strata of the Salt Lake Formation between the Dayton-Oxford and East Dayton-Oxford faults (Fig. 3-11a). Slip on the concave, ESE-dipping listric(?) Clifton Basin fault in Clifton Basin likely produced the NW-dipping limb and SW-plunge of the Clifton Basin anticline. The origin of the SE-dipping limb is uncertain, but a west-dipping listric normal fault is probably responsible. The Clifton Basin anticline was probably truncated at depth during excision of the hanging wall by the Clifton fault (Fig. 3-11a, b).

Excision of the Hanging Wall by the Clifton Fault

The Clifton fault appears to be the youngest of all structures associated with the Bannock detachment system in the study area. In some areas the Clifton fault has moved up, into its hanging wall and excised part of the hanging wall, in a process shown diagrammatically in Fig. 3-2. The Oxford Ridge anticline is the only structure in extension episode 1 that is younger than the Clifton fault (Table 3-3). Excision occurs when a splay of the master detachment fault slices into the upper plate (Lister and Davis, 1989). In this process, the slice of hanging wall becomes part of the footwall of the master detachment (Fig. 3-2). Lister and Davis (1989) explain how excisement and incisement of the upper and lower plates, respectively, occurred in the Whipple Mountains detachment system. As the lower plate (footwall) of the system bows upward, splays of the master detachment fault slice into the upper plate (hanging wall) (i.e. excision) and multiple detachment surfaces are produced (Fig. 3-2). Such a relationship also exists in the Chemehuevi Mountains of southeastern California (John, 1987). There the low-angle, normal Mohave Wash fault lies structurally below the Chemehuevi detachment fault and is interpreted as an early detachment surface (John, 1987).

Several key observations support the interpretation that the Clifton fault is the youngest structure in the detachment system: (1) No low-angle normal faults in the hanging wall of the Clifton fault, as well as many high-angle listric and planar faults, cut the Clifton fault. In several places, they clearly sole into or are truncated by the Clifton fault. For example, map patterns show that the Pocket Basin fault merges northward with the Clifton fault just west of Oxford Ridge (Fig. 3-5, point A). Also, the hanging-wall fault blocks on Oxford Ridge either sole into or are cut by the Clifton fault (Figs. 3-5, 3-9a, b). The Clifton Basin fault is truncated by the Clifton fault; (2) The Clifton fault is very consistent in its geometry across the study area and cuts units with a wide range of dips and dip directions in its hanging wall (Link, 1982a; this study); (3) The Clifton fault cuts every subset of faults (1b and c) and folds associated with extension episode 1. It cuts the ESE-plunging Oxford Peak anticline that is confined to its hanging wall. The Clifton fault was also active during deposition of the upper two members of the Salt Lake Formation in its hanging-wall fault blocks (Third Creek and New Canyon members) (Janecke et al., in press); (4) The Clifton Basin anticline, which most likely formed as a rollover anticline above the listric(?) Clifton Basin fault, is truncated below by the Clifton fault (Fig. 3-12).

CROSS FAULTS (FAULT SET 2)

Overprinting the numerous structures associated with the evolving detachment-fault system is a series of high-angle normal faults. These cut the low-angle normal faults of the Bannock detachment system. There are only a few cross faults exposed in the field area, but several others have been identified west of the study area in the Malad City East quadrangle and to the NE in the south Portneuf Range (Danzl, 1985; Sacks and Platt, 1985; Janecke and Evans, 1999; Evans, in prep.). The cross faults mapped in our study area are exposed in the southern half of the study area (Fig. 3-5).

The clearest crosscutting relationship between a cross fault and a low-angle normal fault is at Clifton Creek (Fig. 3-5, point B) There, a high-angle, E-striking, N-dipping normal fault cuts the low-angle Clifton fault. This fault places Cambrian strata in the hanging wall on Pocatello Formation in the footwall. In turn, this high-angle cross fault is truncated by the younger, N-striking Dayton-Oxford fault.

Another cross fault is exposed ~ 1km southeast of Old Baldy Peak (Fig. 3-5). This fault strikes NE and dips NW (Table 3-1). It is a high-angle normal fault that juxtaposes Cambrian strata in the hanging wall on Brigham Group rocks in the footwall. This fault cuts several hanging-wall structures of the Clifton fault, i.e., a large, antithetic, listric fault and several small, low-angle faults. Like the cross fault at Clifton Creek, this fault is truncated to the east by a younger, N-striking Basin-and-Range fault.

The timing of activity on these faults may be coincident with the location of the mantle plume north of the study area in the Eastern Snake River Plain and, therefore, may represent a brief phase of extension associated with subsidence after the passage of the Yellowstone hotspot (Janecke et al., 2000). If so, subsidence toward the Eastern Snake River Plain produced normal faults as far as 120 km from the margins of the plain, in contrast to McQuarrie and Rodgers (1998) who document more localized deformation.

BASIN-AND-RANGE FAULTS (FAULT SET 3)

The main topographic feature in the study area is the Clifton horst, which is bounded on either side by a major, N-striking, Basin-and-Range normal fault. The Dayton-Oxford fault zone bounds the horst to the east and creates a steep range front in the southern Bannock Range (Fig. 3-16). The fault dips ~60° east and extends more than 30 km N-S (Fig. 3-4, Table 3-1). The fault cuts the older Clifton fault as well as two cross faults near Clifton Creek. The Dayton-Oxford fault is mostly concealed by young Quaternary deposits, but a small anticline in deposits of Lake Bonneville in the northeast of the study area suggests that slip on the fault is quite recent (Fig. 3-5). The Dayton-Oxford fault has accommodated a minimum of 1 km of dip slip (Fig. 3-10a).

The East Dayton-Oxford fault is a N-trending, E-dipping normal fault approximately 3.7 km east of the Clifton horst that is likely a separate strand of the Dayton-Oxford fault (Table 3-1). It merges with the Dayton-Oxford fault ~2.5 km south of the northern edge of the study area and about 3 km south of the study area (Fig. 3-4). It cuts and exposes the Tertiary Salt Lake Formation and Brigham Group rocks and the older Clifton Cemetery fault in its footwall. The East Dayton-Oxford fault is completely concealed by Quaternary deposits and slip of the fault is uncertain, but is likely less than slip on the Dayton-Oxford fault. The two strands of the Dayton-Oxford fault resemble strands of the East Cache fault on the opposite side of Cache Valley in both their spacing, topographic expression, and presence of Salt Lake Formation in the intervening fault block (McCalpin, 1994; Brummer, 1991; Oaks et al., 1999).

The Clifton horst is bounded on the west by the Deep Creek fault, which is a N-striking normal fault that has a west dip of about 55° (Table 3-1). Northward, the fault changes strike from north to northwest in the west-central part of the study area. This fault cuts the Pocket Basin fault and is inferred to cut the Clifton fault. The hanging wall of the fault is the Deep Creek half graben. The fault offsets the Salt Lake Formation and younger Pleistocene-Pliocene(?) deposits in the Clifton horst from similar deposits in the Deep Creek half graben (Fig. 3-5). Slip on the Deep Creek fault is a minimum of 600 m, less than that of its eastern counter-part, the Dayton-Oxford fault. Pleistocene-Pliocene(?) piedmont deposits are preserved only on the west side of the Clifton horst. This suggests that the Deep Creek fault is less active than the Dayton-Oxford fault after deposition of the piedmont deposits.

Several extensional folds are related to the Deep Creek fault. The west limb of the Pocket Basin syncline was likely produced by isostatic rebound of the Deep Creek fault (Figs. 3-5, 3-12). An unnamed NW-trending syncline that folds Tertiary Salt Lake Formation is located in the hanging wall of the Deep Creek fault (Fig 3-5). Several other pairs of synclines and anticlines are prevalent in strata of the Salt Lake Formation in the Deep Creek half graben (Figs. 3-5, 3-

10a). These folds trend roughly parallel to the Deep Creek fault and lesser faults associated with it (Fig. 3-5). Many of the folds in the Deep Creek half graben appear to be controlled by normal faulting associated with the youngest, Basin-and-Range episode of extension.

DISCUSSION

Stretched Pebbles and Metamorphism of the Pocatello Formation

Stretched pebbles in the Pocatello Formation were likely deformed prior to slip on the Bannock detachment system. They do not appear to have been influenced by extension on the Bannock detachment system and exhumation of its footwall during the Tertiary. Metamorphism of the Pocatello Formation is thought to have occurred during the late Mesozoic (Link, 1982a, 1982b). There is no clear evidence to link metamorphism to exhumation of the footwall of the Bannock detachment system or to conclude a Tertiary age.

Summary of Extensional Events

The first major phases of extension in the study area spanned approximately 8 million years (~12 to 4 Ma), from the late Miocene to early Pliocene (Table 3-3, episode 1) (ages from Janecke et al., in press). This phase of extension occurred on low-angle normal faults of the Bannock detachment system. Restoration of the regional cross section (Fig. 3-19a, b) indicates that at least 15 km of WSW-directed extension occurred on the detachment-fault system (approximately 50% extension). Like Janecke and Evans (1999) and Evans (in prep.), we propose that the SW-dipping Valley fault of Sacks and Platt (1985) is the breakaway for the Bannock detachment system. The hanging wall of the detachment system began as a cohesive block during which the Cache Valley Member of the Salt Lake Formation was deposited (Table 3-3, episode 1b). As slip on the fault progressed, the ESE-plunging, Oxford Peak anticline formed. Later the hanging-wall broke-up along listric and planar normal faults both synthetic and antithetic to the master (Clifton) fault (Table 3-3, episode 1c). The Third Creek and New Canyon members of the

Salt Lake Formation were then deposited in the small basins between hanging-wall fault blocks (Table 3-3, episode 1c). Rapid thinning of the crust in the hanging wall resulted in initial doming of the Bannock detachment system by the Oxford Ridge anticline. The Clifton fault excised its hanging wall (Fig. 3-2), but doming continued and folded the Clifton fault (Table 3-3, episode 1d).

Cross faults record a brief phase of N-S to NW-SE-directed extension in the study area following detachment faulting (after 4 Ma) (Table 3-3, episode 2). The Yellowstone hotspot was located approximately 200 km north-northeast of the study area around 4 Ma. Based on migration rates of 30 km/m.y. and the present distribution of seismic zones, the study area would have been in the active belt, Belt II, of the seismic parabola of Pierce and Morgan (1992) around 4 Ma. Activity of the cross faults in the area is coincident with the projected seismic zone around the hotspot, so cross faulting may have been influenced by the hotspot's position.

The third phase of extension began after late Pliocene. It is defined by high-angle normal faults like these of the Basin-and-Range province (Table 3-3, episode 3). E-W extension has been accommodated by N-striking normal faults. Pliocene-Pleistocene(?) gravels were deposited during a brief period of quiescence following the initial uplift of the Clifton horst (Table 3-3, episode 3b). Continued activity of the Basin-and-Range faults brought the Clifton horst to its present topographic elevation (Table 3-3, episode 3c). The area currently resides in the Belt III seismic zone of the Yellowstone hotspot of Pierce and Morgan (1992). Overall, 60% extension was accommodated by all three episodes of extension.

Cache-Pocatello Culmination

The regional extent of the Bannock detachment system coincides well with the regional extent of the Cache-Pocatello culmination of the Sevier fold-and-thrust belt (Janecke et al., in press; Chapter 2 of this thesis) (Figs. 3-3, 3-7). Low-angle faults with attitudes and hanging-wall and footwall rocks similar to those of the Bannock detachment system have been documented north and south of our study area (Trimble, 1976; Link, 1982a, 1982b; Link et al., 1993; Rodgers and Othberg, 1996; Goessel et al., 1999; Kellogg et al., 1999; Oaks et al., 1999; Kruger et al., in press). The Cache-Pocatello culmination extends from Cache Valley in Utah at latitude 41° 40' N north to the Snake River Plain at 42° 00' N, a distance of at least 135 km (Figs. 3-3, 3-7). Similarly, the Bannock detachment fault has a proposed N-S extent of over 134 km from northern Utah to the Pocatello area and northward to the east Snake River Plain.

We believe that the Bannock detachment system collapsed the Cache-Pocatello culmination during the late Miocene to Pliocene. Extensional collapse of thrust-related culminations has been documented in several locations along the Sevier fold-and-thrust belt in the eastern Great Basin (Schirmer, 1988; Bryant, 1990; DeCelles et al., 1995; Constenius, 1996; Coogen and DeCelles, 1996; Yonkee, 1997). The Bannock detachment system is very similar to the Sevier Desert detachment system, which collapsed a culmination with similar structural characteristics as the Cache-Pocatello culmination (DeCelles et al., 1995). Both culminations are broad anticlines with low topographic relief developed above an upper flat east of a major footwall ramp in the Sevier fold-and-thrust belt (this study; DeCelles et al., 1995; Coogan and DeCelles, 1996).

Collapse of the Cache-Pocatello culmination likely started more than 50 million years after it formed. Eastward transport by the Paris-Willard thrust likely ended by the Late Cretaceous (DeCelles, 1994) and collapse by the Bannock detachment system began ~12 Ma. Thermal weakening by the Yellowstone hotspot may have triggered the collapse of the Cache-Pocatello culmination.

The Bannock detachment system accommodated about 50% extension in southeast Idaho and Basin-and-Range extension added about another 10%. The Cache-Pocatello culmination presently has an E-W extent of about 54 km. Before extension, its E-W extent was about 33 km.

Origin of the Oxford Peak Anticline

The Oxford Peak anticline is an ESE-trending fold that formed in a cohesive hanging wall of the detachment system probably during the initial translation across the fault system. The axial trace of this anticline trends 106° and plunges 24° ESE. It projects roughly toward the intersection zone between the N-striking East Cache fault and the NNW-striking Valley fault (Figs. 3-4, 3-20a). The origin of the Oxford Peak anticline is uncertain, but it must be an extensional fold. The subcrop map (Fig. 3-7) shows little evidence for the anticline prior to extension. Also, the ESE-plunge of the anticline is orthogonal to N- and NNW-trending Sevier-age structures.

There are three possible mechanisms that we considered to explain the origin of this fold: (1) An increase in slip south of the fold axis, on the Bannock detachment system, could have rotated the southern limb with respect to a fixed northern limb (Fig. 3-20b). However, this seems unlikely because the required slip gradients and amount of slip predicted for this area of the detachment system on the East Cache fault are too high; (2) Three-dimensional antithetic shear can produce a transverse fold like this above concave-up lateral bends in listric normal faults (Medwedeff and Krantz, 2002). This three-dimensional kinematic model predicts that beds in the hanging wall will dip toward the master normal fault (Fig. 3-20c). The change in strike of the breakaway of the detachment system from NW to N would produce an ENE-plunging anticline (Fig. 3-20c). Limbs of the Oxford Peak anticline are oblique to the master fault, however (Fig. 3-20a), so this model is an incomplete solution; (3) A third possibility is that a large NW-dipping normal fault, southeast of the study area, could have produced the SE-dipping limb of the anticline, perhaps as a large rollover monocline (Fig. 3-20d). However, the only such fault documented in the central or southern Cache Valley area is on the northwest side of Little Mountain southeast of Preston Idaho (Stanley, 1972) and is too short and distant to produce the fold. We believe that the three-dimensional antithetic shear model fits best with our observations, and may be responsible for producing the Oxford Peak anticline. Gradients in slip along the Bannock detachment system may have rotated the limbs of the anticline further.

Origin of the Oxford Ridge Anticline

The Oxford Ridge anticline is a very large, laterally extensive, extensional fold defined by folded foliation in the Pocatello Formation and by folded strands of the Bannock detachment system. The exposures of the east and west limbs of the anticline extend ~8 km east and ~4 km west of the hinge, perpendicular to the fold hinge. At its north end, the Oxford Ridge anticline plunges beneath Marsh Valley.

One of two process may have produced this structure: (1) folding due to isostatic rebound of the crust, or (2) folding due to rollover in the hanging wall of a younger, WSW-dipping, listric normal fault. Fault-drag and fault-propagation folding typically produce smaller folds, and are unlikely to have produced the large Oxford Ridge anticline (Janecke et al., 1998). We believe that the Oxford Ridge anticline likely formed due to isostatic rebound of the footwall of the Clifton detachment fault because there is no laterally continuous WSW-dipping, listric normal fault to the east that could have produced the fold as a rollover anticline.

Several observations suggest that the Oxford Ridge anticline formed during detachment faulting, not during the younger Basin-and-Range phase of deformation. The anticline was produced after faulting of the detachment system began, but the axis of the anticline is parallel to the strike of hanging-wall faults that sole into the master detachment fault and parallel to the strike of the master fault. Thus its geometry is closely linked to the detachment system and not to younger Basin-and-Range faults which mostly strike ~30° clockwise from the Oxford Ridge anticline. With continuous, rapid slip on the detachment system and the break-up of the hanging wall along normal faults, the decrease in overburden thickness allowed the footwall to dome upward and produce the Oxford Ridge anticline (e.g., Spencer, 1985).

Excision of the hanging wall by the Clifton Fault

The process of excision may explain the young age of the Clifton fault relative to all other structures associated with the detachment system, except the Oxford Ridge anticline. We believe that the Clifton fault is the last active splay of the master detachment fault in the Bannock detachment system and that the low-angle normal faults structurally below the Clifton fault are older splays of the detachment system.

A minimum of 5 km of slip occurred on the Clifton fault in order to remove and erode the lens of rock between the original and final position of the fault. Excision thinned the hanging wall of the Clifton fault.

Evidence for Slip at Low Angles on the Clifton Fault

Several lines of evidence in the study area suggest that the master fault, the Clifton fault, as well as other older low-angle normal faults of the Bannock detachment system, were active at low angles of dip and probably originated at low angles. The most convincing evidence for slip at low angles is crosscutting and geometric relationships that show little or no eastward rotation of the master fault or its hanging-wall structures (except on the back-tilted east limb of the Oxford Ridge anticline). Such tilting is predicted by the 'domino' model (Proffett, 1977; Gans and Miller, 1983).

Relationships at Clifton basin indicate slip at a low angle. The Clifton Basin anticline is a double rollover extensional fold created by slip on the listric(?) Clifton Basin fault and the low-angle Clifton Cemetery fault or a similar, now eroded structure. Both of these faults sole into or are cut by the master detachment, the Clifton fault (Figs. 3-10, 3-11). If these faults and this fold are reconstructed to reflect an original steep dip (>45°) of the Clifton fault, the Clifton Basin fault would have been active at a sub-horizontal dip and the Clifton Basin anticline would have been a west-vergent overturned fold. Overturned extensional folds, although possible, are rare (Janecke et al., 1998). The geometries created by this reconstruction are very unlikely. The reasonable conclusion is that the Clifton fault was active at a low angle.

The Pocket Basin fault, which is a hanging-wall imbricate to the detachment system, is sub-parallel to the synextensional deposits of the Salt Lake Formation in its hanging wall. The fault and its hanging wall have been folded to gentle west dips above most of the fault. These low cutoff angles indicate that this part of the Pocket Basin fault was initially sub-horizontal and has since been tilted westward. The 'domino' model predicts an eastward rotation of fault blocks for this area and moderate to steep original dips of faults. The geometries in the Pocket Basin hanging wall are inconsistent with the domino model.

At Rattlesnake Ridge just south of the study area, similar relationships as those in the Pocket Basin area indicate formation and slip on the Clifton fault at a low angle. There, the Salt Lake Formation in the hanging wall of the Clifton fault dips to the west and strikes sub-parallel to the west-dipping Clifton fault. The Neoproterozoic Pocatello Formation in the footwall of the Clifton fault also has a west-dipping fabric parallel to the fault. Locally a horse of early Paleozoic rocks is caught along the Clifton fault. The Salt Lake Formation and the Clifton fault exhibit a flat-on-flat geometry indicating an original low-angle dip of the Clifton fault and subsequent gentle westward tilting in the hanging wall of a younger Basin-and-Range fault of extension episode 3. Again, the 'domino' model predicts eastward tilting of the fault.

The lack of thick deposits of Tertiary Salt Lake Formation in the hanging wall of the Clifton fault indicates that the fault and it hanging-wall faults had original low-angle dips. If the Clifton fault originated as a steep, high-angle fault, a thick sequence of basin fill deposits, >8.5 km, should be preserved next to the fault, to match the > 10 km of slip across the fault. The cumulative thickness of the Salt Lake Formation is only 2-3 km in the area (Danzl, 1982; Sacks and Platt, 1985; Goessel et al., 1999; Goessel, 1999; Evans, in prep). The Clifton fault does not display the geometries predicted by the 'domino' model.

Another reason the Bannock detachment system fits a low-angle, normal-fault model rather than the 'domino' model, is the lack of evidence of a major fault block to the east in Cache Valley. The Clifton fault in the Clifton horst is cut by the younger Dayton-Oxford fault and therefore is down-dropped beneath Cache Valley. If the originally SW-dipping, low-angle Clifton fault was rotated to a low angle through time, then the large, down-faulted, east-rotated fault block should surface or be detectable to the east in Cache Valley. Gravity and seismic data indicate that there is very little basin fill in northern Cache Valley (Peterson and Oriel, 1970; Bankey and Kleinkopf, 1984; Evans and Oaks, 1996). Therefore, the presence of a major rotated fault block should be detectable. None has been found. There is no evidence in Cache Valley to support the rotation of the entire detachment fault system from a steeper dip. The only evidence for the continuation of the low-angle fault system beneath Cache Valley is at Little Mountain, northwest of Preston, Idaho (Figs. 3-4, 3-19). There a low-angle, gently, ESE-dipping normal fault crops out with Brigham Group rocks in the hanging wall and Pocatello Formation in the footwall. This fault is interpreted to be a continuation of the master low-angle fault, the Clifton fault (Link and LeFebre, 1983), on the east limb of the Oxford Ridge anticline. The Brigham Group rocks there are interpreted as part of a rider block in the hanging wall of the master detachment. The domino model predicts Salt Lake Formation in the hanging wall of the Clifton fault at that location.

The Bannock detachment system resembles detachments that bound metamorphic core complexes, many of which are known to have formed and been active at low angles (John, 1987; Lister and Davis, 1989; Miller and John, 1999). (1) Faults in the hanging wall of the Clifton fault are closely spaced, listric and planar normal faults that either sole into the Clifton fault or are cut by it. These relationships are unlike those of areas with rotated fault blocks (Proffett, 1977); (2) In some areas the low-angle hanging-wall fault blocks of the Clifton fault bound very thin slices of rock (e.g., those just west of Oxford Ridge). Such patterns are more typical of detachment-fault terranes than of rotated fault-block terranes; (3) The Oxford Ridge anticline, which folds the Clifton fault and its footwall, is similar to the ‘domed’ core of metamorphic core complexes (Spencer, 1985; Lister and Davis, 1989; Yin and Dunn, 1992); (4) A mylonitic fabric similar to those associated with the core of a metamorphic core complex is interpreted along the Clifton fault in the Pocatello Formation north of the study area (Link, 1982a, 1982b), but most of the footwall is non-mylonitic; (5) Part of the hanging wall of the Clifton fault was excised by the Clifton fault. This process is only documented in metamorphic core complex terranes (John, 1987; Lister and Davis, 1989); (6) A possible breakaway fault to the detachment system, common in metamorphic core complexes, is present east of the study area. Sacks and Platt (1985) showed that the Valley fault, a listric normal fault in the Portneuf Range to the east, is flat at about 4 km below the surface. Because the Valley fault becomes a rather flat and shallow normal fault at depth east of Cache Valley (Sacks and Platt, 1985) and because there is a lack of large fault blocks in Cache Valley, we suggest that the Valley fault is the breakaway for the Bannock detachment system; (7) It is difficult to accumulate large amounts of slip on steeply to moderately dipping normal faults. The Clifton fault slipped at least 15 km with excision taken into account. Such large offsets have not been described on moderately dipping normal faults. ‘Domino’-style extension alone is unlikely to produce a fault system with as much slip as the Bannock detachment system.

CONCLUSIONS

The Bannock detachment system is a regionally extensive system (>134 km N-S) that accommodated up to 50% extension. The master fault of the system is structurally the highest, most laterally continuous, and youngest of the low-angle faults in the system. It was active at a low angle, probably formed at a low angle, and likely excised part of its own hanging wall, a process which has only been documented in major extensional systems (John, 1987; Davis and Lister, 1988; Lister and Davis, 1989). Extension on this detachment system produced two large, laterally extensive anticlines, the Oxford Peak anticline and the Oxford Ridge anticline, at right angles to one another. The latter we interpret to be the proto-core of the detachment system similar to the core of a metamorphic core complex. All of the characteristics described for the Bannock detachment system have also been documented in detachment fault systems that were active at low angles. Few of the structural characteristics described for the Bannock detachment system resemble those of an extensional system with domino-style, fault-block rotation without a basal detachment. We interpret the Bannock detachment system as a major extensional system of Miocene-Pliocene age.

The Bannock detachment system is very similar to the Sevier Desert detachment system. The Bannock detachment system lies in the same structural position as the Sevier Desert detachment system, has a similar vergence, and original geometry. Both systems collapsed culminations at the western edge of the Sevier fold-and-thrust belt.

Structural events recorded in the study area, from oldest to youngest, are as follows: (1) Formation of the Cache-Pocatello culmination above and east of the Malad ramp of the Paris-Willard thrust, accompanied and followed by erosion of Mesozoic and upper Paleozoic rocks down to the Cambrian-Ordovician stratigraphic level. Some metamorphism in the Pocatello Formation may date from this time period; (2) Development of localized graben; (3) Initiation of the Bannock detachment and initial collapse of the Cache-Pocatello culmination; (4) Development of the ESE-plunging Oxford Peak anticline in the hanging wall of the Bannock detachment system; (5) Break-up of the hanging wall of the Bannock detachment system on SW-dipping faults sub-parallel to the master fault of the detachment system. (6) Excision of the hanging wall by the Clifton fault and doming of the footwall of the Bannock detachment system after excision; older strands of the detachment are preserved in the footwall of the Clifton fault; (7) Development of generally E-W-striking, normal cross faults that cut the detachment system; (8) Initiation of high-angle Basin-and-Range faults and the initial uplift of the Clifton horst; (9)

Deposition of Pliocene-Pleistocene(?) piedmont gravels during a brief period of quiescence of the Basin-and-Range faults (Janecke et al., in press; Chapter 2 of this thesis); (10) Continued extension and uplift of the Basin-and-Range horsts to their present topographic expression.

Deformation during the first episode of extension on the Bannock detachment system was three-dimensional in time and space and was very complex. Superposition of structures from the two extensional events that followed greatly added to the three-dimensional complexity of the field area. Further work, especially in the footwall of the detachment system, is needed to fully characterize this surprisingly young and surprisingly large extensional system that formed within the northeast part of the Basin-and-Range province astride a culmination in the dormant thrust belt of the Sevier Orogeny.

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TABLE 3-1. DESCRIPTION OF MAJOR NORMAL FAULTS

Fault	Strike and dip at surface of fault plane	Episode of extension	Geometry used in cross sections	Length (km)	Amount of dip-slip displacement (km)	Comments
Clifton	Variable 339º, 17º WSW in N 331º, 17º SW in M 349°, 9° WSW in M 347º, 21º WSW in S	1b, 1d	Listric east of Cache Valley; folded-planar in study area	>42	15.1 minimum, with excision taken into account (~10.1 without excision) Cross section B-B'	Master low-angle fault. Youngest of detachment faults.
Pocket Basin	350º, 18º WSW	1c	Listric, folded	>7	>2.5 Cross section D-D'	Hanging-wall imbricate to Clifton fault. Likely soles into Clifton fault.
Clifton Cemetery	320º, 21ºSW	1c	Listric	>5	>2.8 Cross section A-A'	Hanging-wall imbricate to Clifton fault. Soles into or is cut by Clifton fault.
Clifton Basin	010º, ~30ºE	1c	Listric	1.6	>4.7 Cross section C-C'	Antithetic fault to Clifton fault. Cut by Clifton fault at Clifton basin. Produced NW-dipping limb of Clifton basin anticline.
Low-angle fault S of Weston Peak	021º, ~5°E	1c	Listric	2.5	~0.4 Cross section A-A'	Antithetic to and soles into Clifton fault. Possible southern extension of Clifton Basin fault.
Cross fault at Clifton Creek	081°, ~60° N	2	Planar	1	<0.3	Cuts Clifton fault, but is cut by Dayton-Oxford fault.
Deep Creek	Variable 324º, ~55ºSW in N 358ºN, ~53ºW in S	3	Listric to slightly listric?	>16, in study area	>0.6	Bounds Clifton horst on the west
Dayton-Oxford	350-359º, ~60ºE	3	Planar	>30	>1.0	Bounds Clifton horst on the east. Cuts Clifton fault and Clifton Cemetery fault.
East Dayton-Oxford	324-005º, ~60ºE	3	Planar	>21	uncertain	Fault strand synthetic to Dayton-Oxford fault

TABLE 3-2. DESCRIPTION OF MAJOR EXTENSIONAL FOLDS

Fold	Orientation (trend and plunge of fold axis)	Bisecting surface (strike and dip)	Interlimb angle (gentle, open, closed)	Number of bedding (or foliation*) attitudes used to define fold	Inferred Fold type	Geometry of fold	Associated normal fault(s)	Fault set fold belongs to	Angle between trend of fold axis and strike of fault
Oxford Ridge anticline	345°, 0° *(NNW)*	345°, 90°	131, open	35*	Isostatic	Longitudinal	Clifton	1	Sub-parallel to Clifton fault (6º)
Oxford Peak anticline	106°, 24° *(ESE)*	103°, 84°W	149, gentle	60	Three-dimensional, fault bend	Transverse	Bannock detachment system	1b	Oblique (55º)
Clifton Basin anticline	217°, 15° *(SW)*	211°, 65°W	105, open	16	Double rollover	?	Clifton Cemetery and Clifton basin faults	1c	Approximately perpendicular to Clifton Cemetery fault (78º) and sub-parallel to Clifton Basin fault
Pocket Basin syncline	313°, 16° *(NW)*	135°, 83°W	143, gentle	8	Uncertain	Longitudinal	Deep Creek faults	3a	Parallel to Deep Creek fault (11º)

*Foliation

TABLE 3-3. SUMMARY OF TECTONIC EVENTS IN NORTHERN CACHE VALLEY

Designation	Structural developments	Sedimentary deposit or unconformity associated with this event (from Janecke et al., in press)	Age (from Janecke et al., in press)	Specific Structures (those marked * are outside of study area)
Episode 1 Fault set 1a	Small localized rift basins filled mostly with conglomerate	Skyline Member of the Salt Lake Formation, Collinston Conglomerate, and Red Conglomerate of Sacks and Platt (1985)	Older than 10.27±0.07 Ma ($^{40}Ar/^{39}Ar$ date) and probably older than 10.94 Ma deposits in Junction Hills (Oaks, 2000)	• *Beaver Dam fault (Goessel et al., 1999) • *unnamed fault in west half of Deep Creek half graben (Evans, in prep)
Episode 1 Fault set 1b	Initiation of the Bannock detachment system and gravitational collapse of Cache-Pocatello culmination. Development of Oxford Peak anticline	Cache Valley Member of Salt Lake Formation	Before 10.27±0.07 Ma until about 10.13± 0.3 Ma	• Proto-Clifton fault
Episode 1 Fault set 1c	Breakup of the hanging wall of the Bannock detachment system along SW-dipping normal faults. Development of Clifton basin anticline	Third Creek and New Canyon members of the Salt Lake Formation Third Creek Member is the first to contain clasts from then Neoproterozoic Brigham Group	Before 9.6 ± 0.2 Ma until after 5.1 to 4.4 Ma (possibly until about 2 to 3 Ma)	Light blue faults on Fig. 3: • Pocket Basin fault; • Clifton Cemetery fault; • Faults in hanging wall of Clifton fault in excision area • Clifton Basin fault • Low-angle fault under Weston and Baldy peaks
Episode 1 Fault set 1d	Development of Oxford Ridge anticline. Excision of hanging wall by the Clifton fault.	Third Creek and New Canyon members of the Salt Lake Formation?		• Clifton fault (dark blue fault on Fig. 3) • Oxford Ridge anticline
Episode 2 Fault set 2 (Hot spot-related flexure and faulting?)	Formation of cross faults with E, NE and NW strikes	Little or no sedimentary record?	Mostly after episode 1d and before episode 3a; after 5.1 to 4.4 Ma	• Green faults in Fig. 3 • Older cross faults of S. Clarkston Mountain and Junction Hills (Goessel et al., 1999; Oaks, 2000) • 10 NE-striking normal faults in S. Cache Valley (Smith, 1997; Oaks et al., 1999)
Episode 3a Fault set 3 (small-magnitude extension)	Modern Basin-and-Range topography dismembers the older rift basins along N-striking normal faults; Clifton Horst, Malad Range, and Clarkston Mountain are uplifted. Development of Pocket basin syncline?	Erosion and removal of most of the Salt Lake Formation and Wasatch Formation, plus lower Paleozoic and Neoproterozoic rocks from the Clifton Horst	After 5.1 to 4.4 Ma	Red faults in Fig 3: • Deep Creek fault • Dayton-Oxford fault • East Dayton-Oxford fault • Small faults in Deep Creek half-graben
Episode 3b Fault set 3	Tectonic quiescence and near burial of the Clifton Horst by piedmont gravel deposits which contain clasts of Neoproterozoic Pocatello Formation	Deposition of piedmont gravel (QTG)	After 5.1 to 4.4 Ma	
Episode 3c Fault set 3	Renewed uplift of the Clifton Horst	Erosion of piedmont gravel; external drainage to the Great Salt Lake was established by this time, thus lowering the base level	After deposition of piedmont gravel; after early to mid Pleistocene	Red faults on Fig. 3 • Dayton-Oxford fault • East Dayton-Oxford fault • Deep Creek fault?

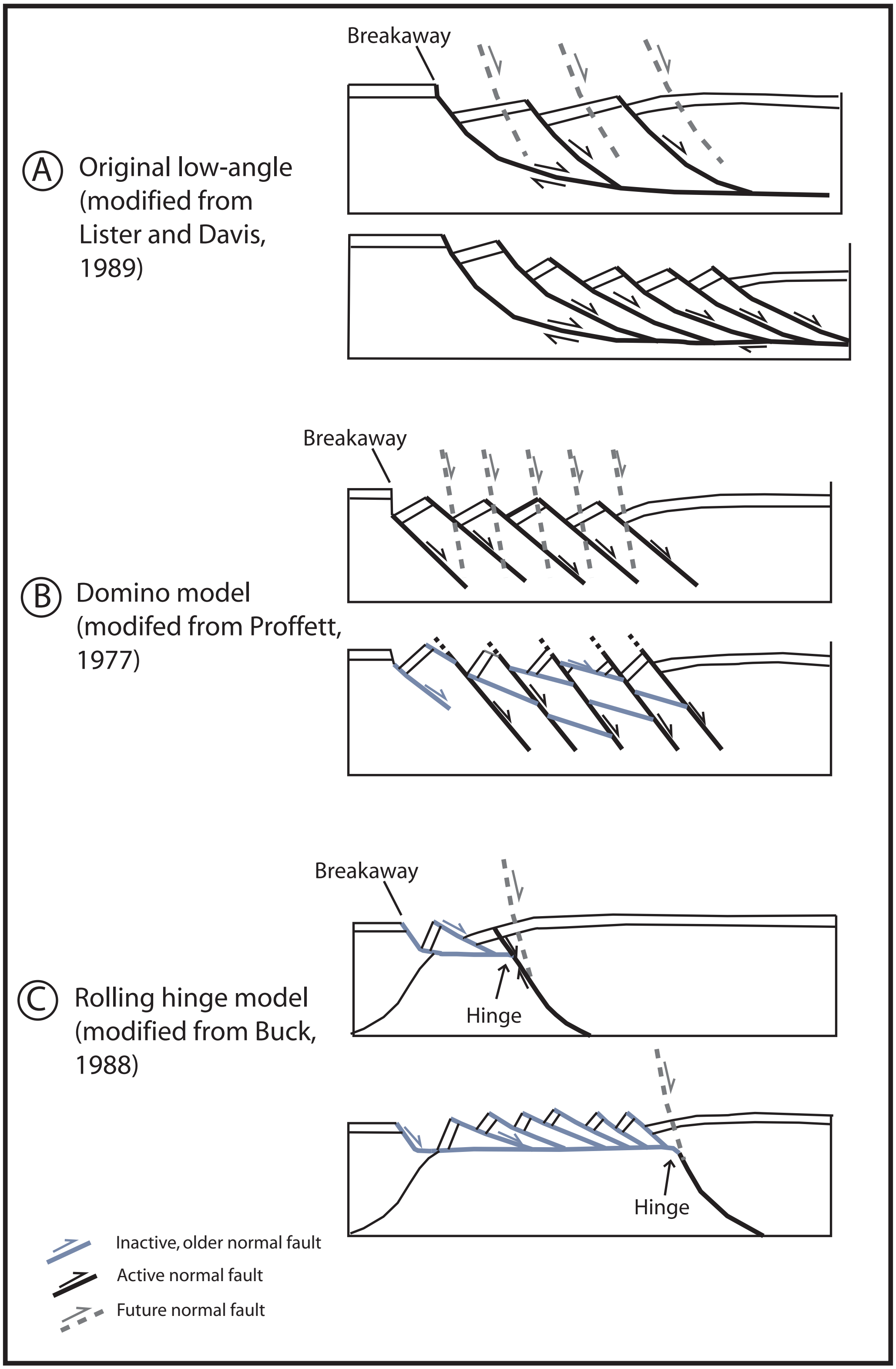


Figure 3-1. Schematic diagrams showing three models of extension in detachment fault systems.

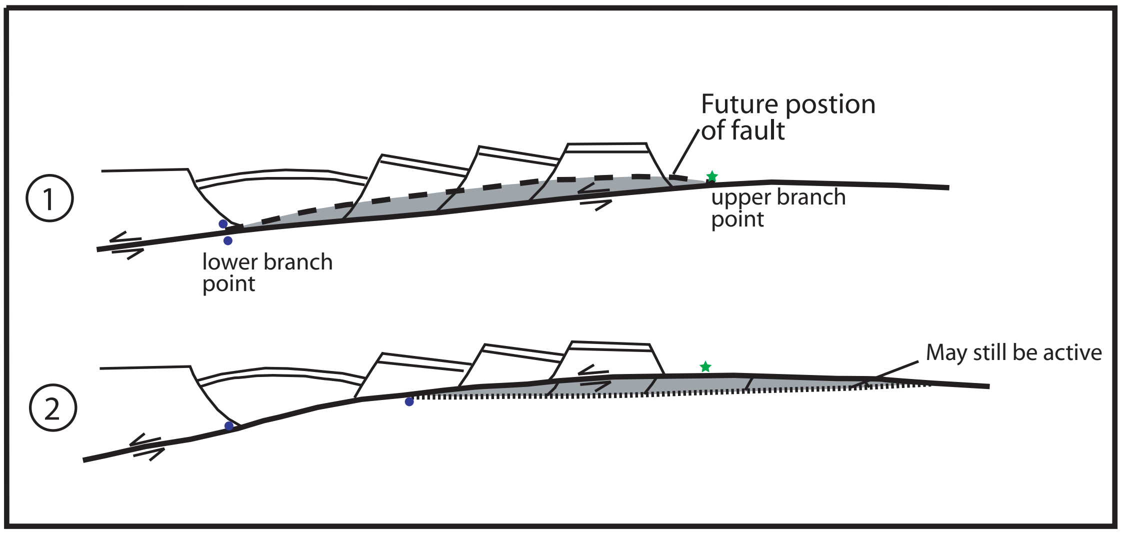


Figure 3-2. Schematic diagram to show excision process through time. Note that the lower branch point in the footwall must slip past the upper branch point in the hanging wall in order to remove the excised lens of rock (shaded).

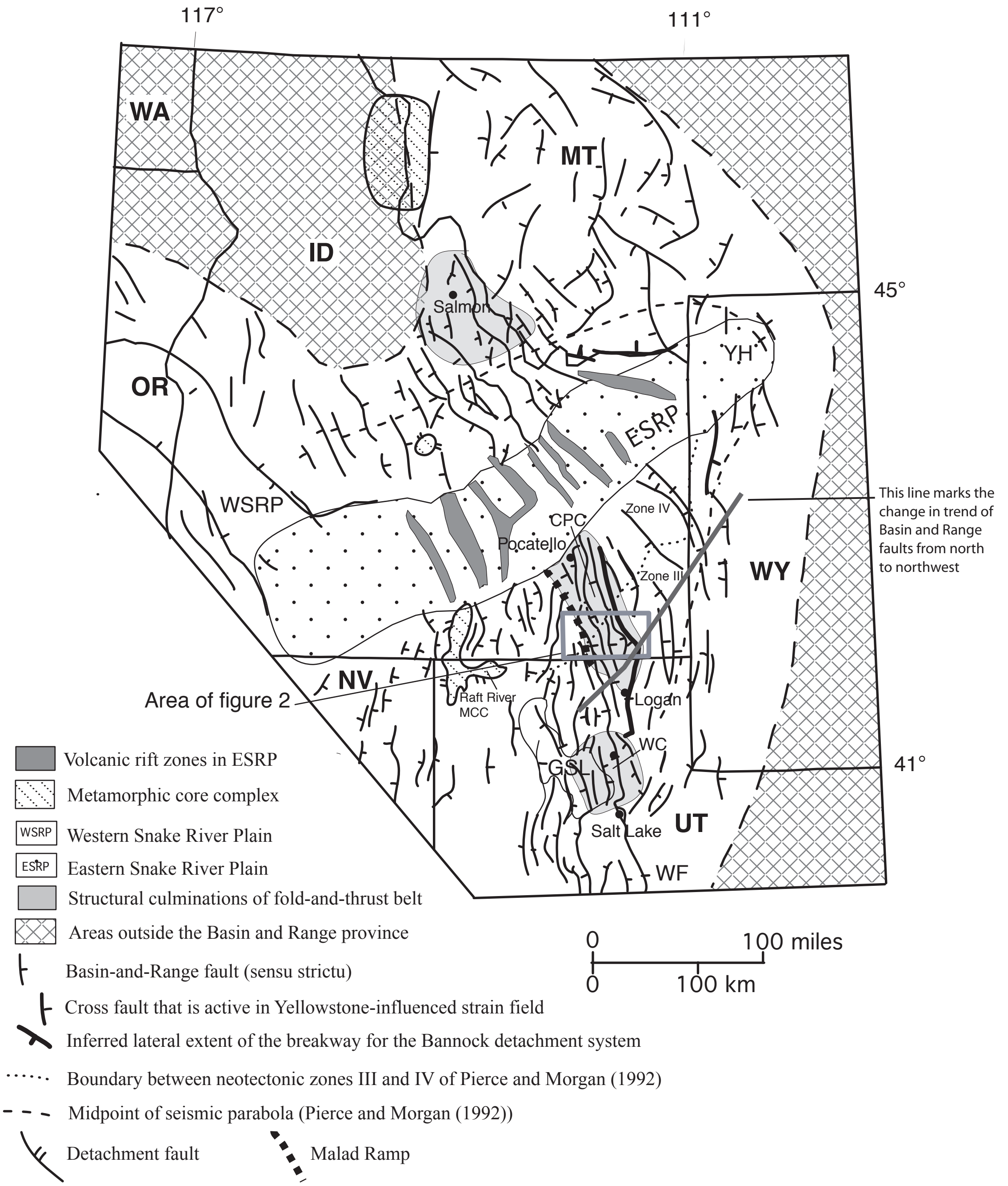


Figure 3-3. Regional map showing the major active normal faults adjacent to the Eastern Snake River Plain (ESRP), the north-northwest-trending Cache-Pocatello culmination (CPC), and selected other features. Note that the inferred Bannock detachment system essentially coincides with the Cache-Pocatello culmination and has the opposite vergence as the partly coeval Raft River metmorphic core complex (Wells et al., 2000). The areally less extensive Wasatch culmination 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), Janecke et al. (2001) and this study.

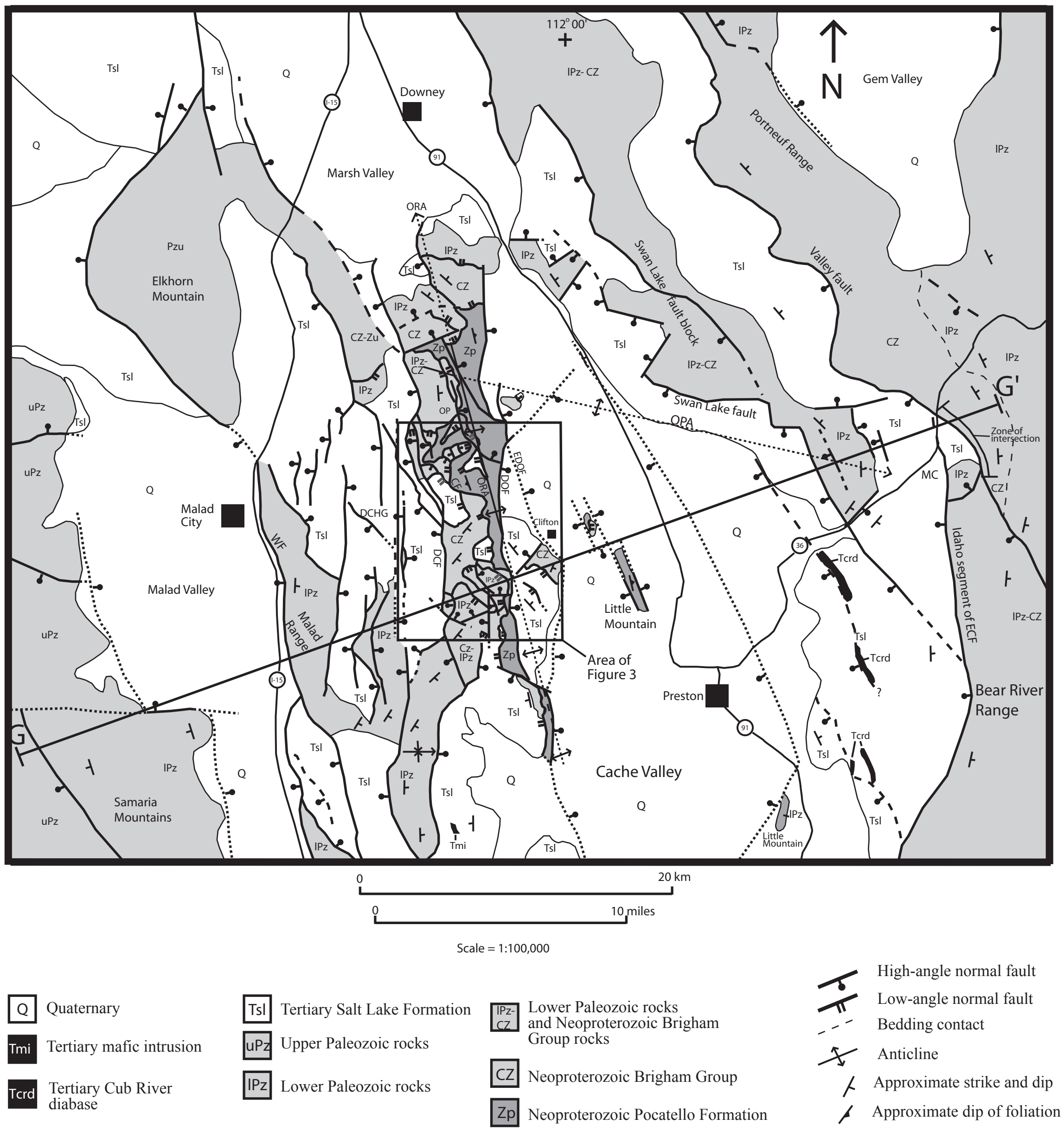


Figure 3-4. Simplified geologic map of the northern Cache Valley area in Idaho showing Basin and Range normal faults and low-angle normal faults of the Bannock detachment system. The Clifton horst is between the Deep Creek fault and the Dayton-Oxford fault. Abbreviations are : CF= Clifton fault, DCF= Deep Creek fault, DCHG = Deep Creek half graben, DOF = Dayton-Oxford fault, ECF = East Cache Fault, EDOF = East Dayton-Oxford fault, MC = Mink Creek, OP = Oxford Peak, OPA = Oxford Peak anticline, ORA = Oxford Ridge anticline. G-G' is a geologic cross section in Figure 10. 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 et al. (in press), Janecke, unpublished mapping, and this study.

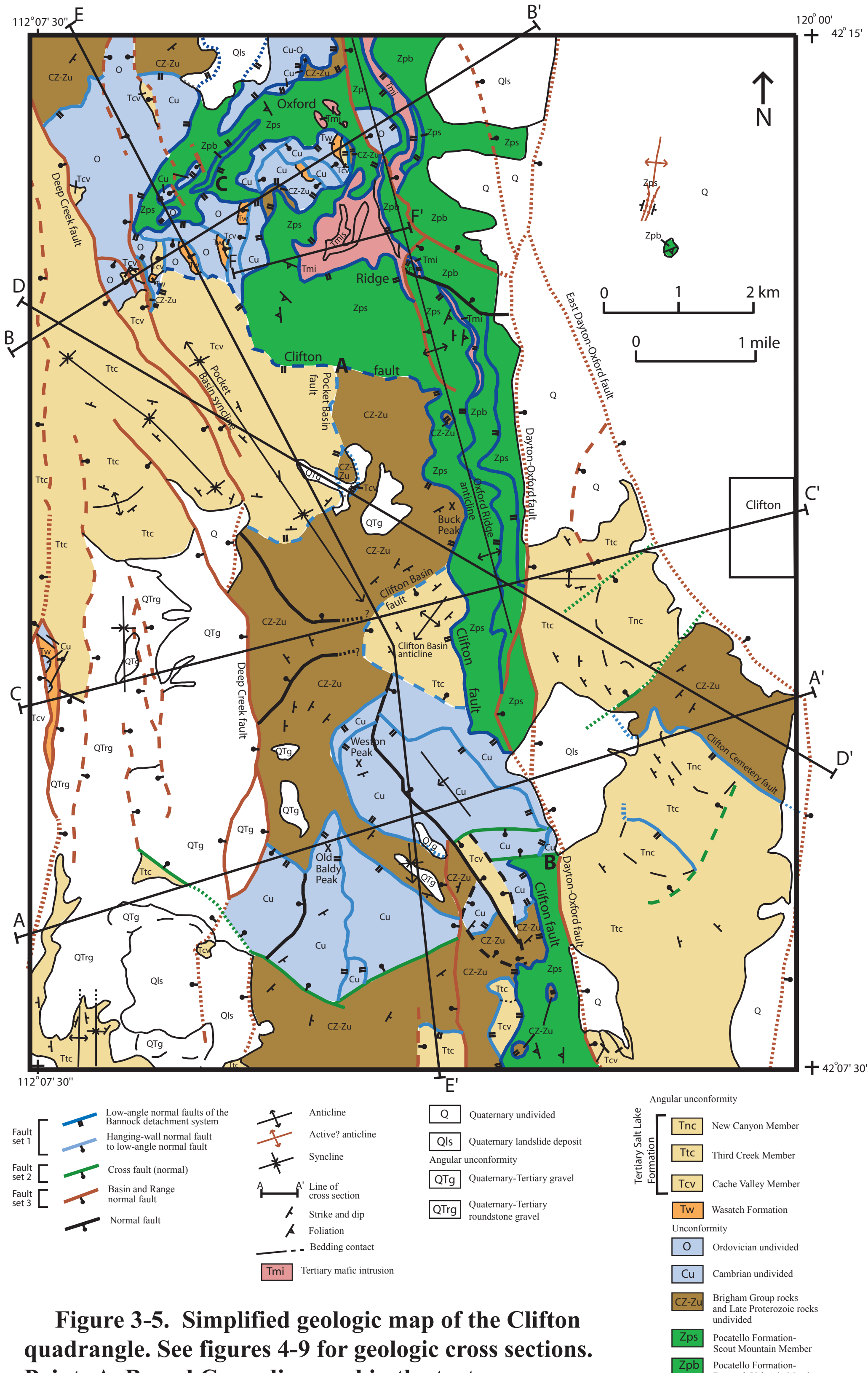


Figure 3-5. Simplified geologic map of the Clifton quadrangle. See figures 4-9 for geologic cross sections. Points A, B, and C are discussed in the text.

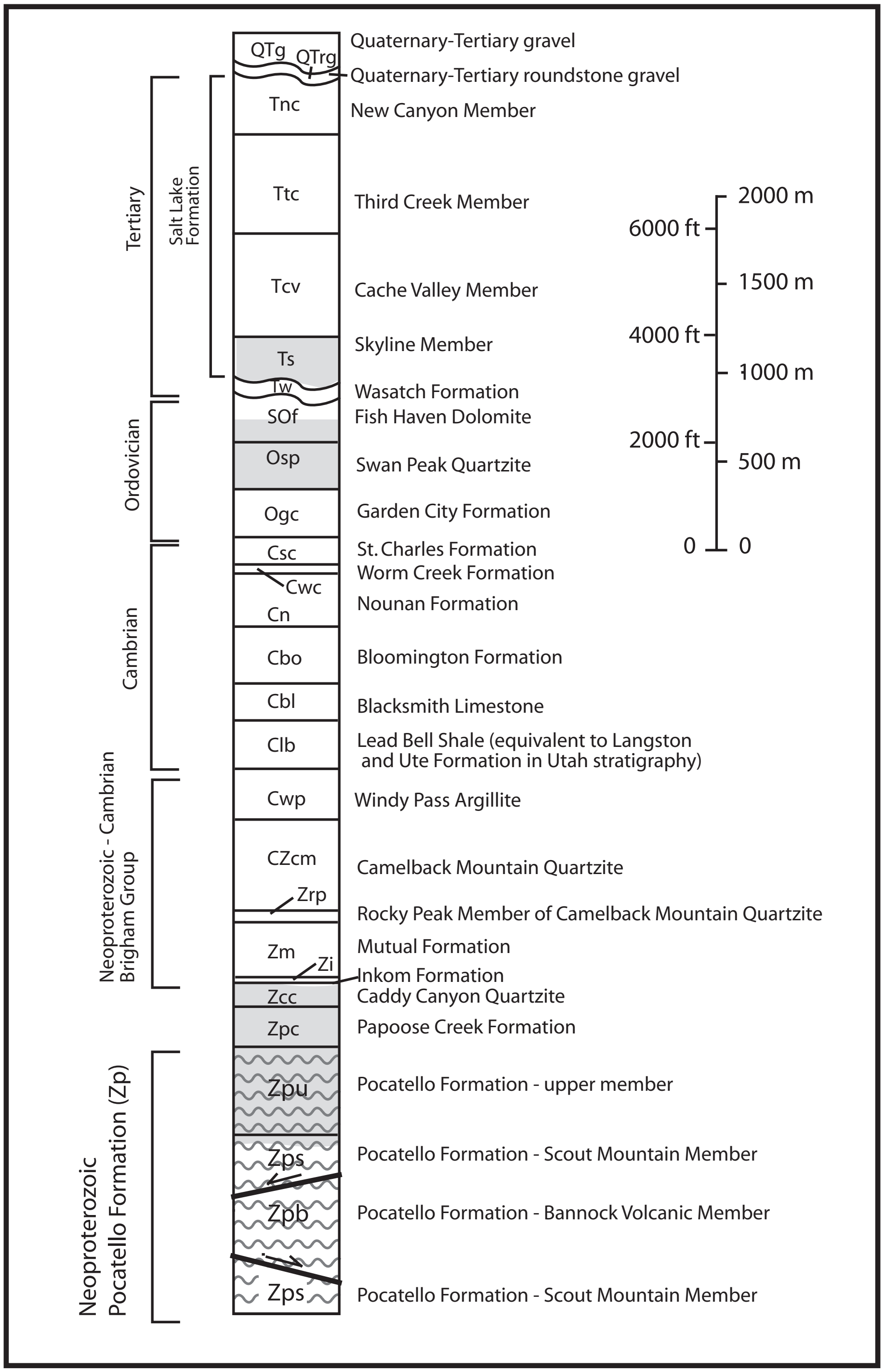


Figure 3-6. Generalized stratigraphic column of strata exposed in the Clifton quadrangle. Thicknesses are minimum estimates. Shaded areas indicate strata that is missing, not exposed, or faulted-out. Modified from Evans (in prep); original figure by Link (1982b).

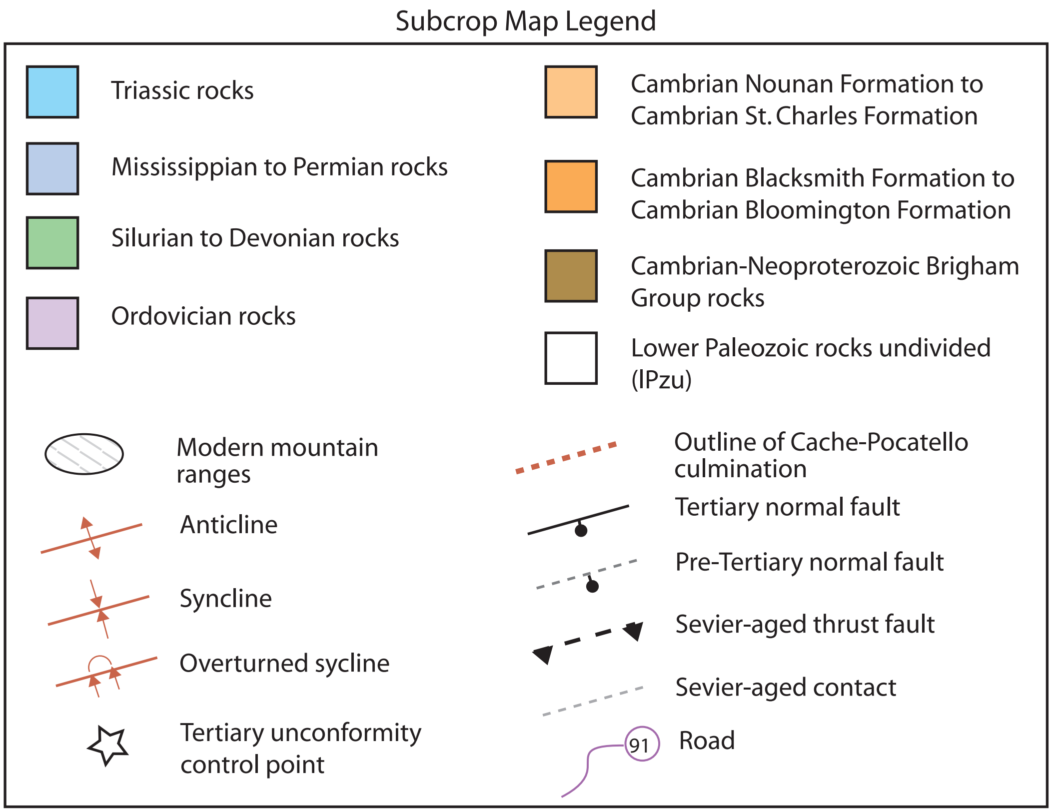


Figure 3-7. Subcrop map, overlain on modern topography, showing the location and age of rocks beneath the Tertiary unconformity. Also shown are Mesozoic folds and thrusts of the Sevier Orogeny, the location of the Oxford Peak anticline (OPA), and the location of the Cache Pocatello culmination. Updated and expanded from Rodgers and Janecke, 1992. Compiled from Mansfield (1927), Prammani (1957), Murdock (1961), Oriel and Platt (1968), Armstrong (1969), Raymond (1971), Willard (1972), Shearer (1975), Mayer (1979), Rember and Bennett (1979), Link (1982a), Sacks (1984), Danzl (1985), Sacks and Platt (1985), Oriel et al. (1991), Lowe and Galloway (1993), Dover (1995), Evans et al. (1996), Smith (1997), Goessel (1999), Goessel et al. (1999), Link and Stanford (1999 and maps cited there in), Oaks et al. (1999), Crane (2000), Biek et al. (2000), Oaks (2000), Biek et al. (2001), Pope et al. (2001), DeVecchio (2002), Evans (in prep), and this study. In mountain ranges, the age of the youngest exposed rock unit was also used to construct the subcrop map. The effects on younger normal faults were mostly ignored. Abbreviations: DOF = Dayton-Oxford fault, EDOF = East Dayton-Oxford fault, WF = Wasatch fault.

Figure 3-7 continued.

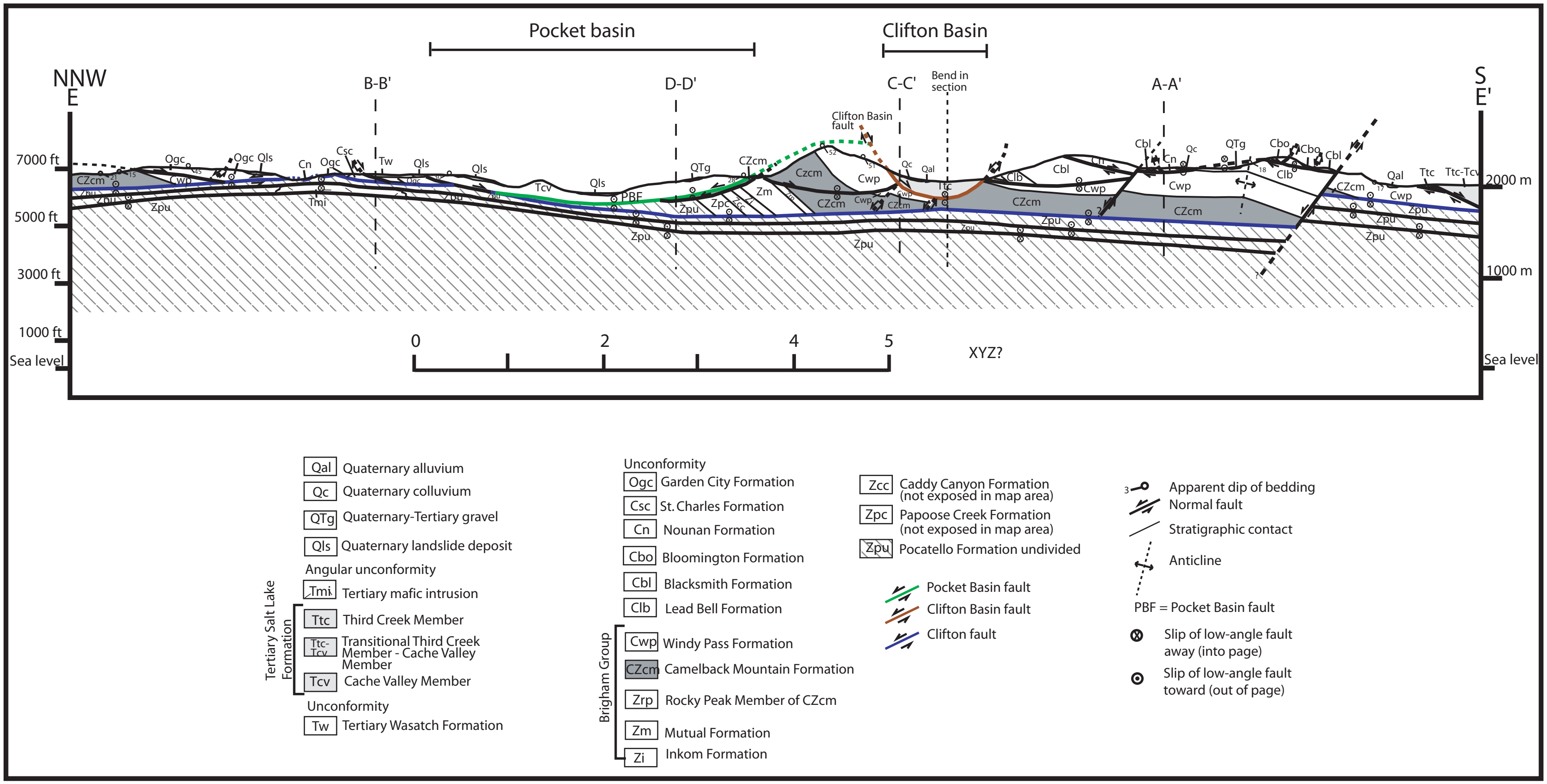


Figure 3-8. Geologic cross section E-E' along the Clifton horst. See Figure 3 for location of section.

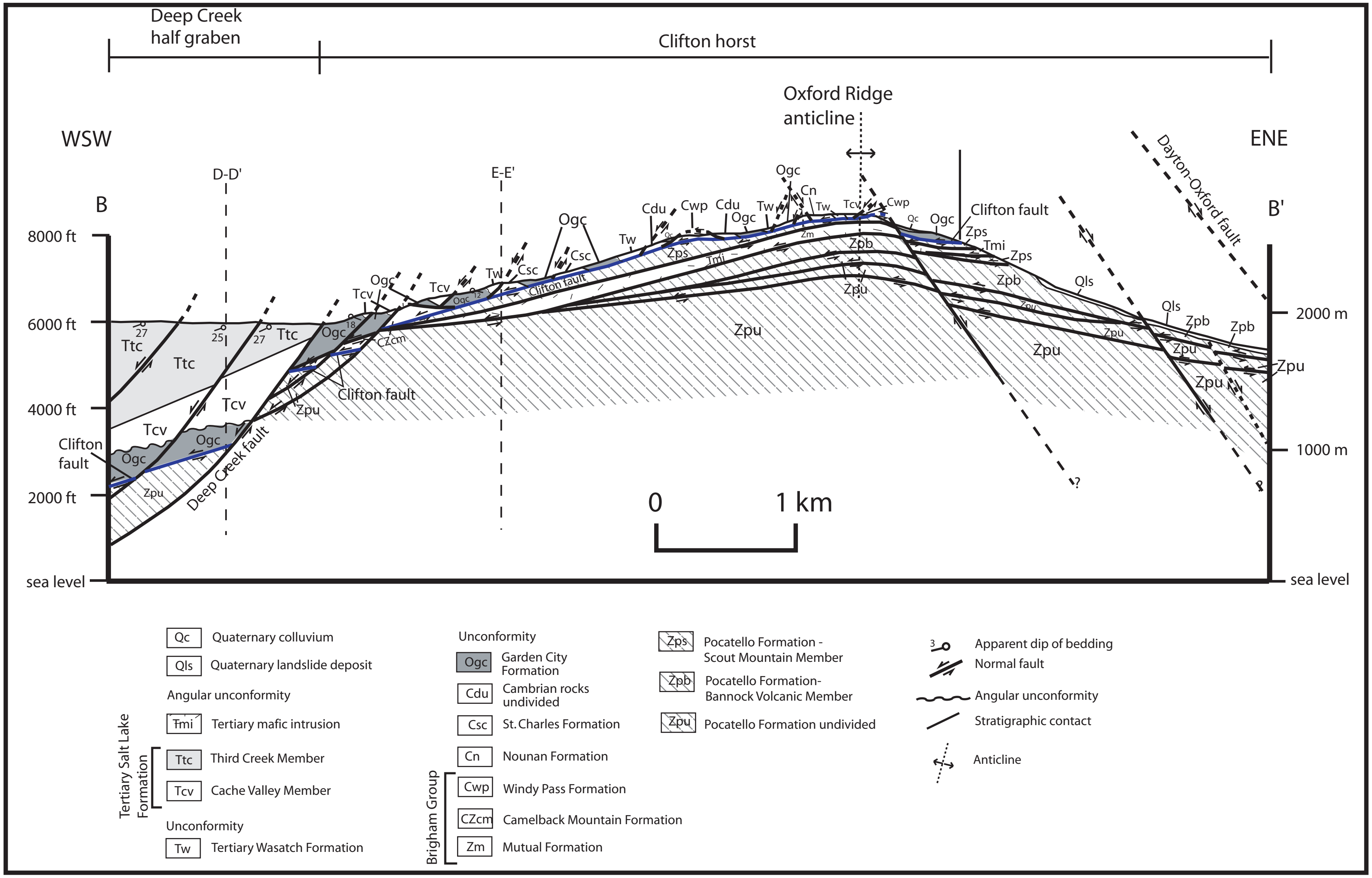


Figure 3-9a. Geologic cross section B-B' across the Clifton horst in the study area. Relationships beneath Clifton fault are uncertain. See Figure 3 for location of this section.

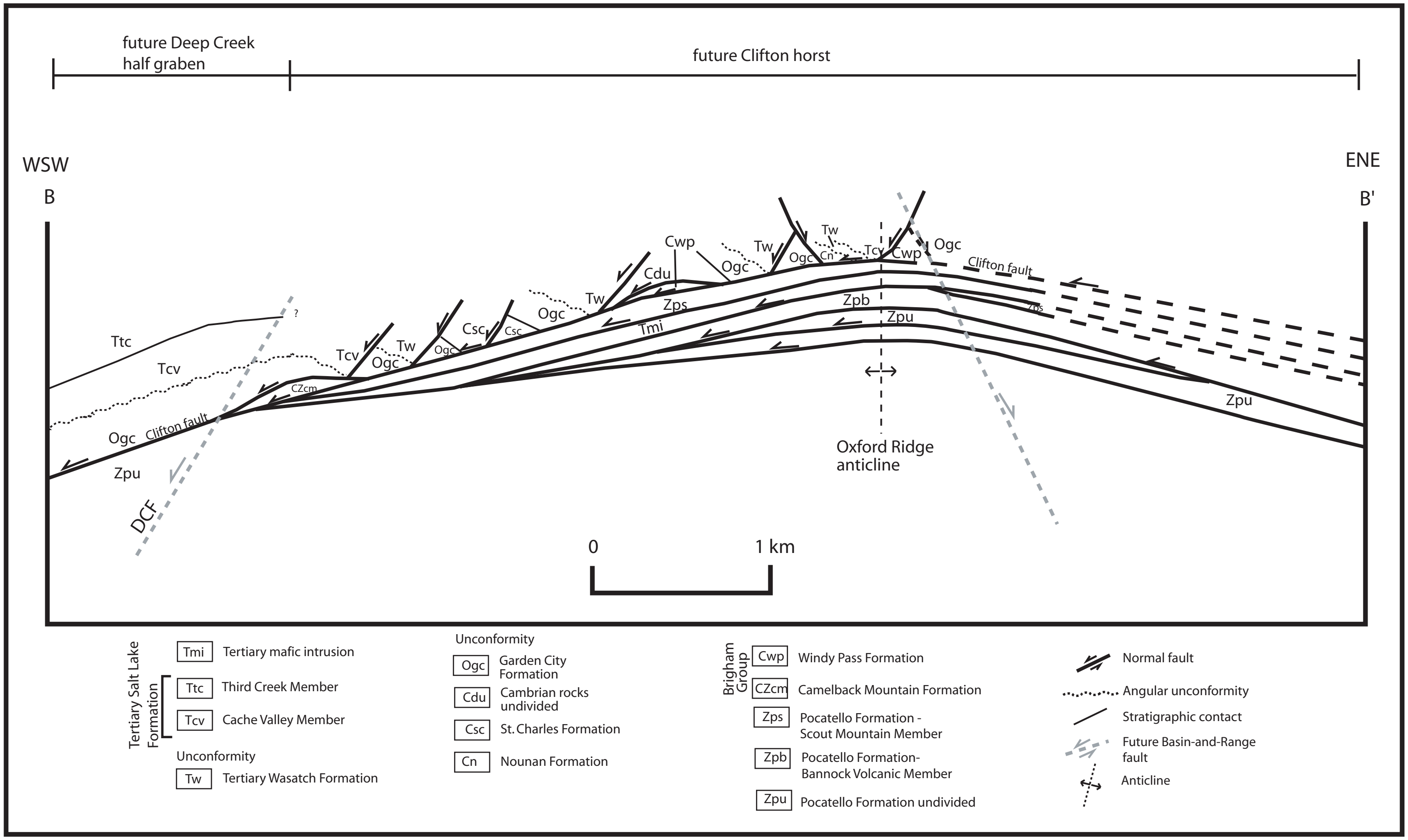


Figure 3-9b. Reconstruction of cross section B-B' before Basin-and-Range extension. Approximate future positions of Basin-and-Range topography and faults are shown. Abbreviations: DCF = Deep Creek fault.

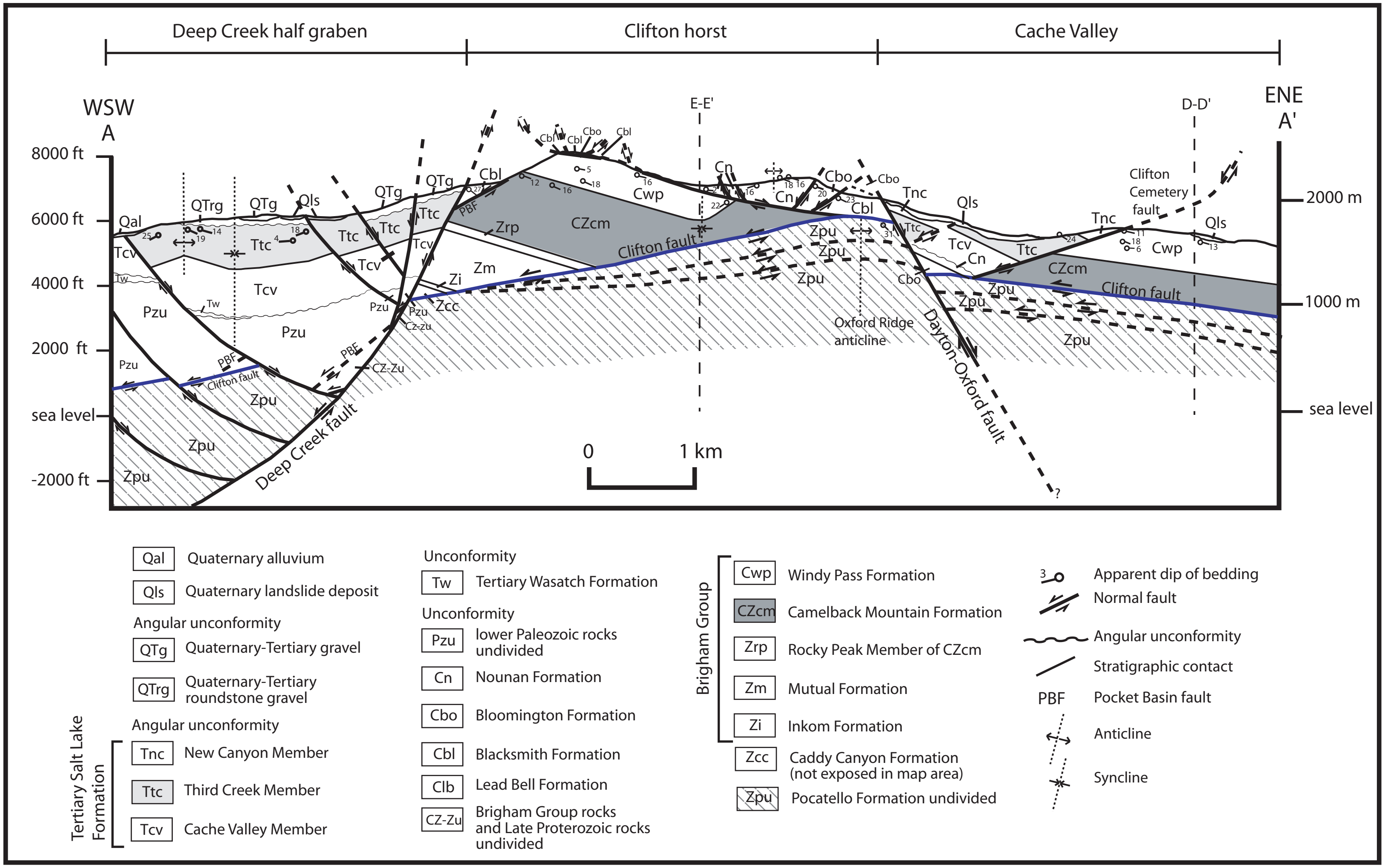


Figure 3-10a. Geologic cross section A-A' across the Clifton horst in the study area. Relationships beneath Clifton fault are uncertain. See Figure 3 for location of this section.

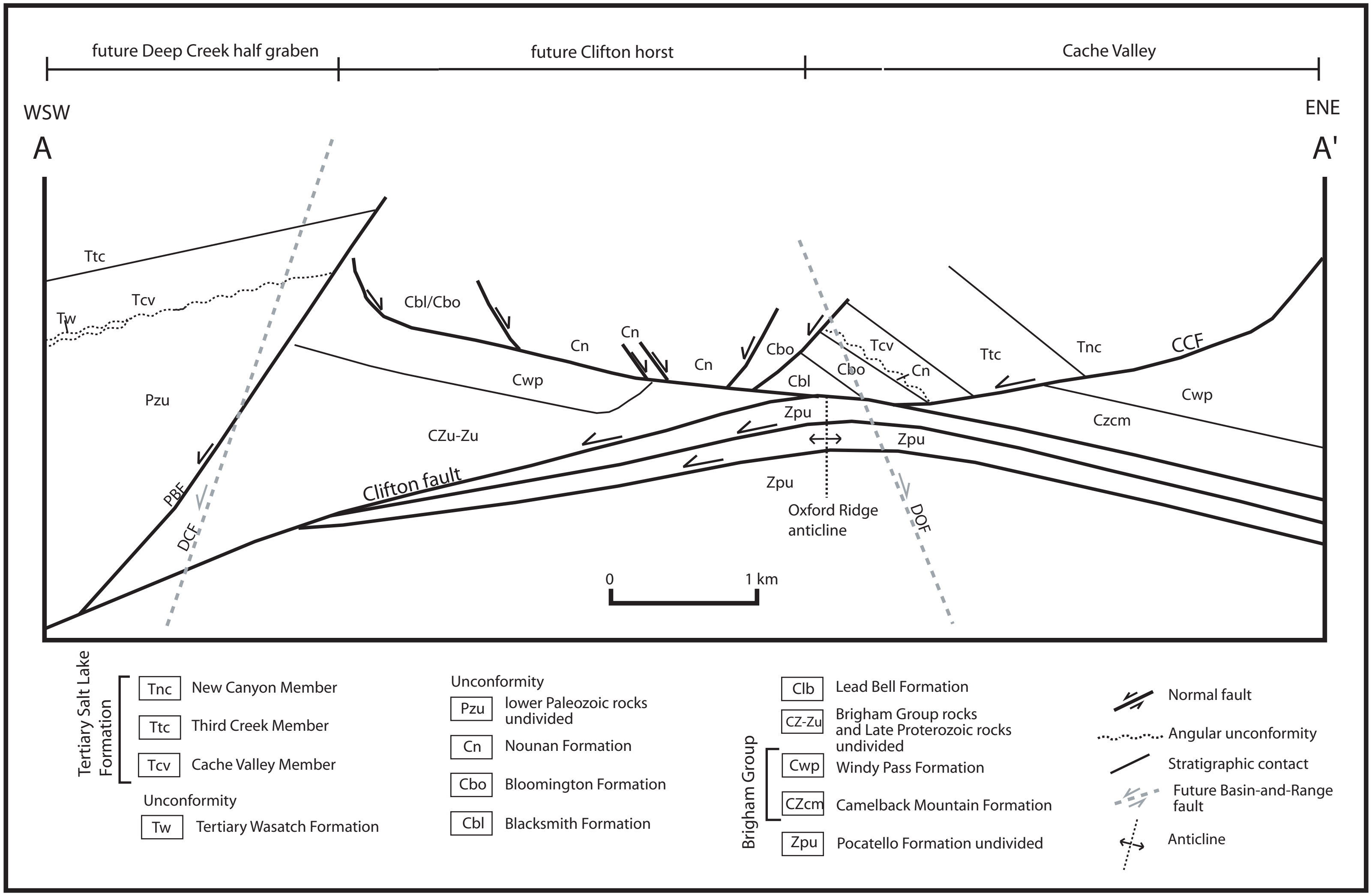


Figure 3-10b. Reconstruction of cross section A-A' before Basin-and-Range extension. Approximate future positions of Basin-and-Range topography and faults are shown. Abbreviations: CCF = Clifton Cemetery fault, DCF = Deep Creek fault, DOF = Dayton Oxford fault, PBF = Pocket Basin fault.

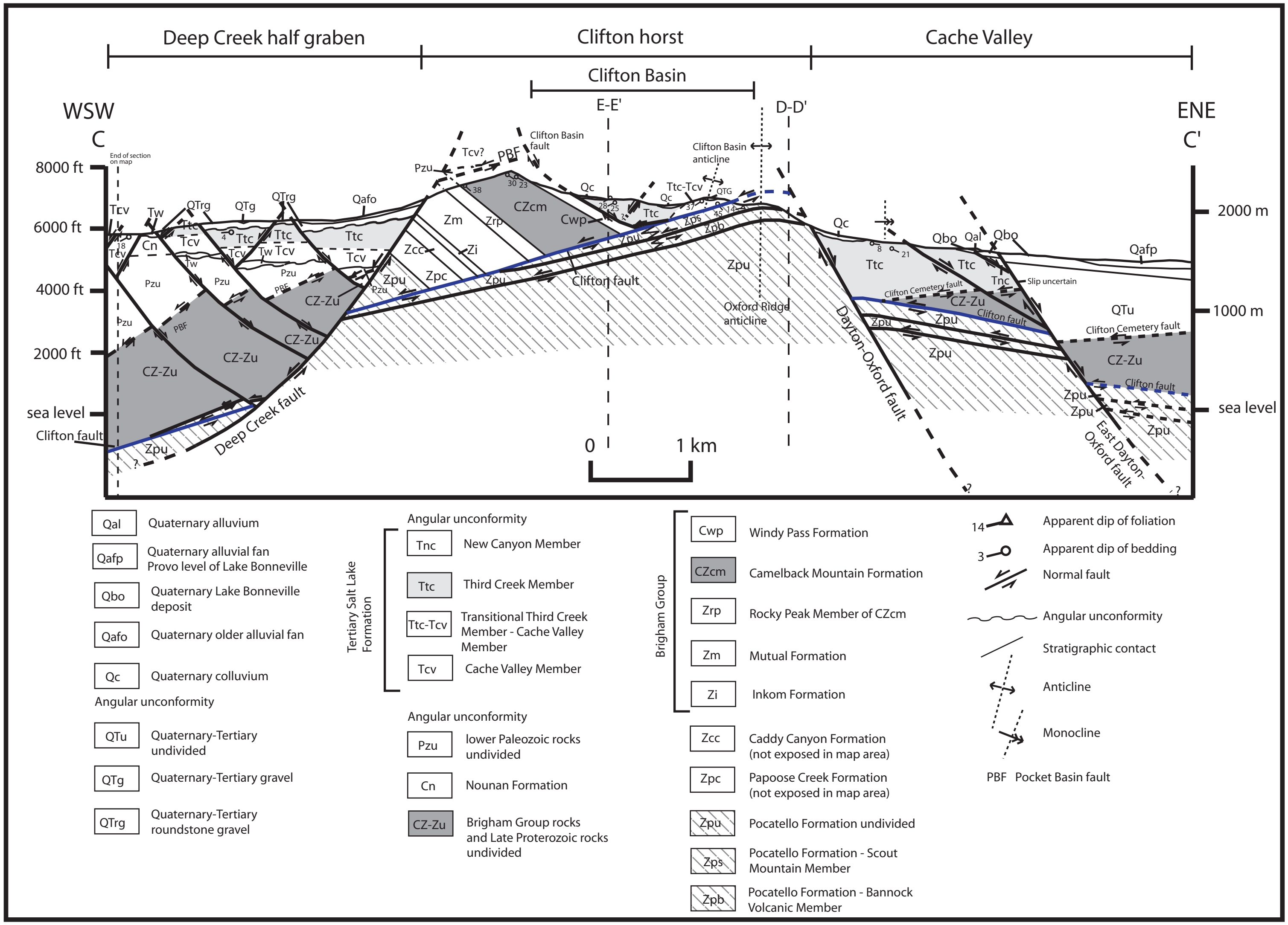


Figure 3-11a. Geologic cross section C-C' across across Clifton horst in the study area. Relationships beneath the Clifton fault and east of the East Dayton-Oxford fault are uncertain. See Figure 3 for location of section.

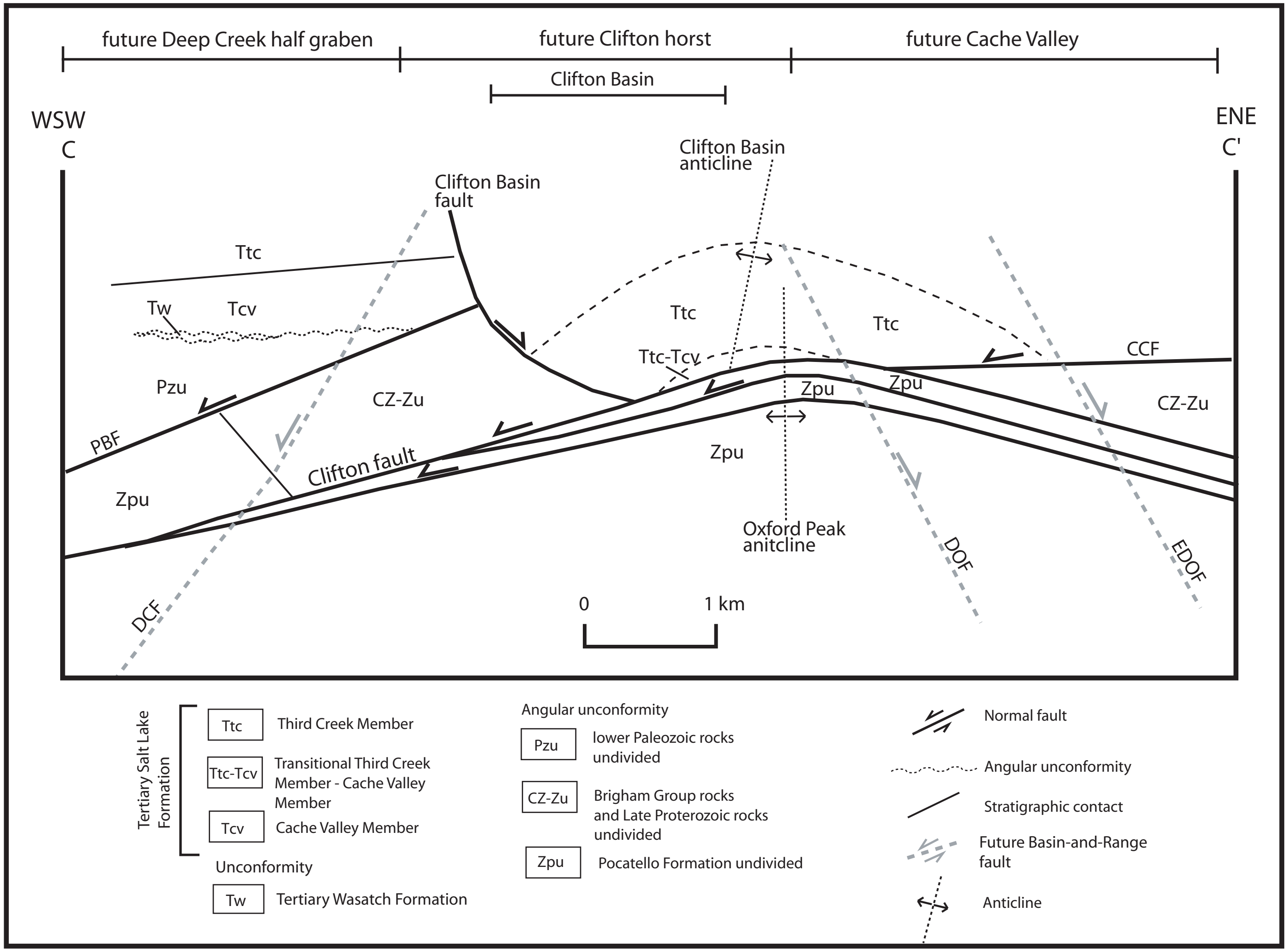


Figure 3-11b. Reconstruction of cross section C-C' before Basin-and-Range extension. Approximate future positions of Basin-and-Range topography and faults are shown. Abbreviations: CCF = Clifton Cemetery fault, DCF = Deep Creek fault, DOF = Dayton-Oxford fault, EDOF = East Dayton-Oxford fault, PBF = Pocket Basin fault.

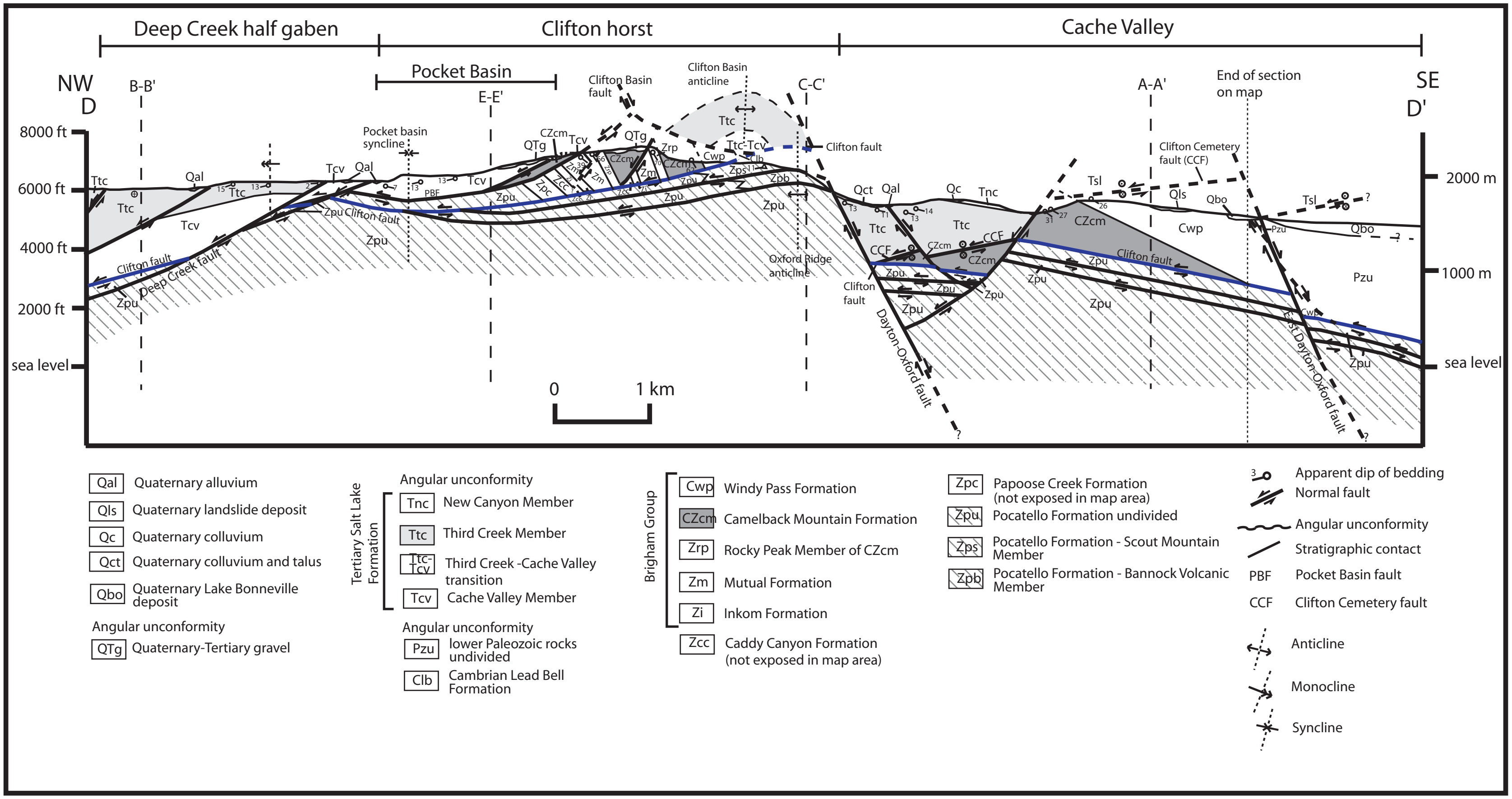


Figure 3-12. Geologic cross section D-D' across the Clifton horst in the study area. See Figure 3 for location of section.

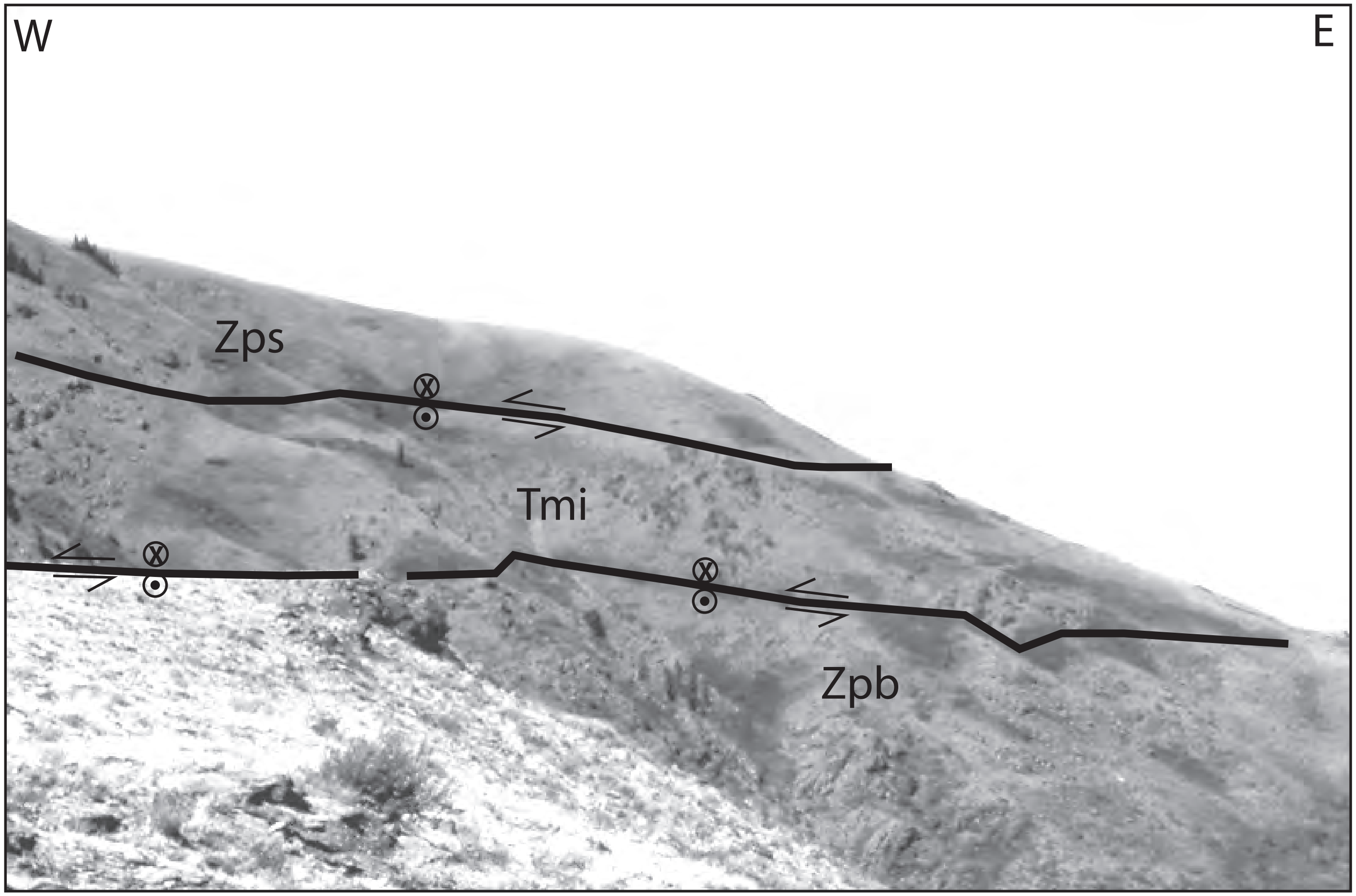
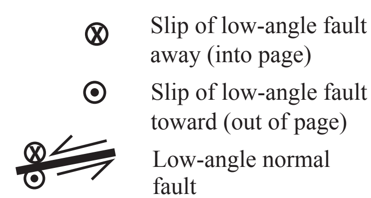


Figure 3-13. View looking north at east face of Oxford Ridge. The low-angle normal faults gently dip east. Abbreviations: Zps = Scout Mountain Member of Pocatello Formation, Zpb = Bannock Volcanic Member of Pocatello Formation, Tmi = Tertiary mafic intrusion.



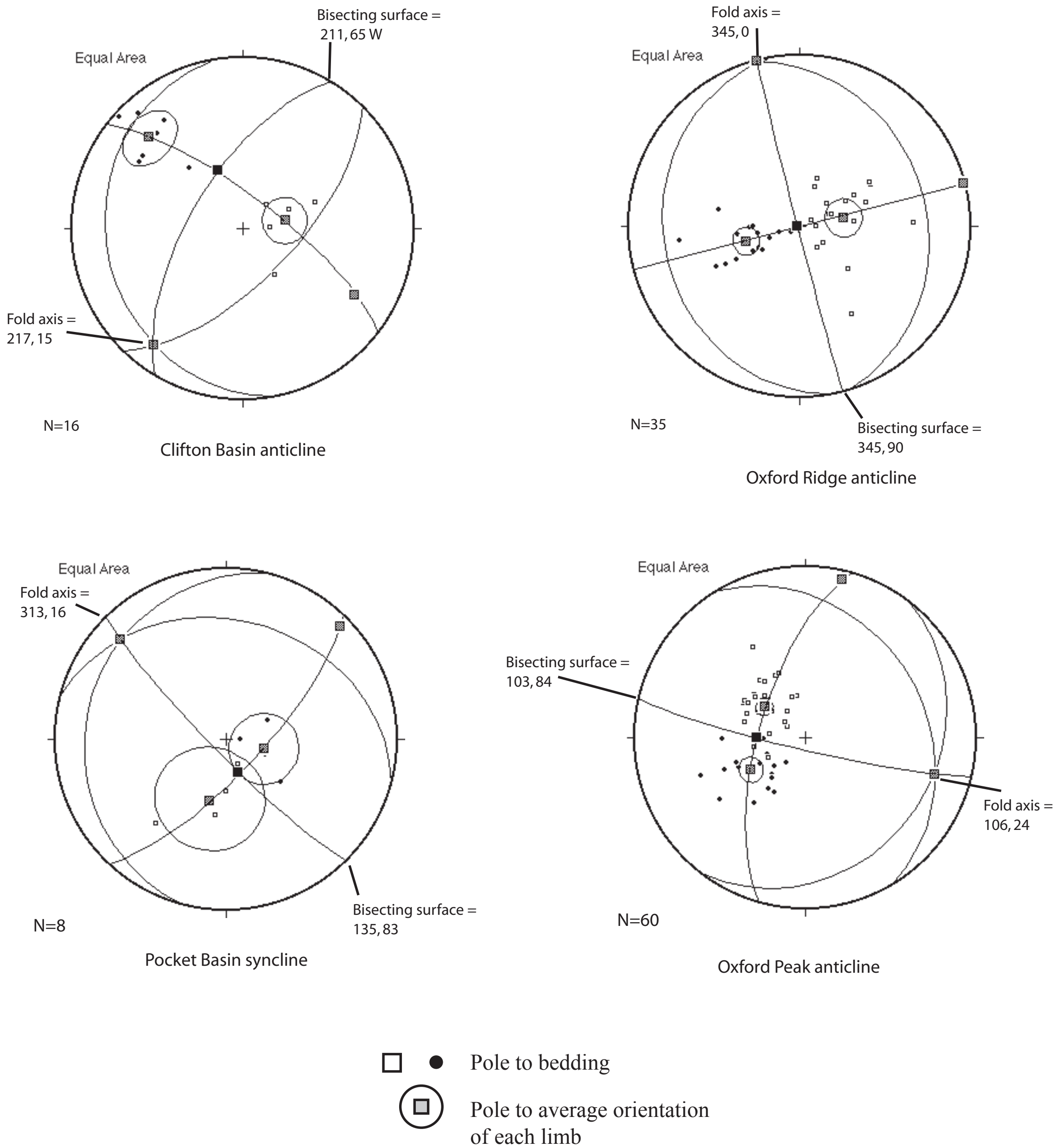


Figure 3-14. Stereograms of poles to bedding (foliation for the Oxford Ridge anticline) for major extensional folds associated with the Bannock detachment system and Basin-and-Range faulting in the study area. See Figure 3 for locations of folds. Mean poles to bedding and foliation were calculated for individual limbs.

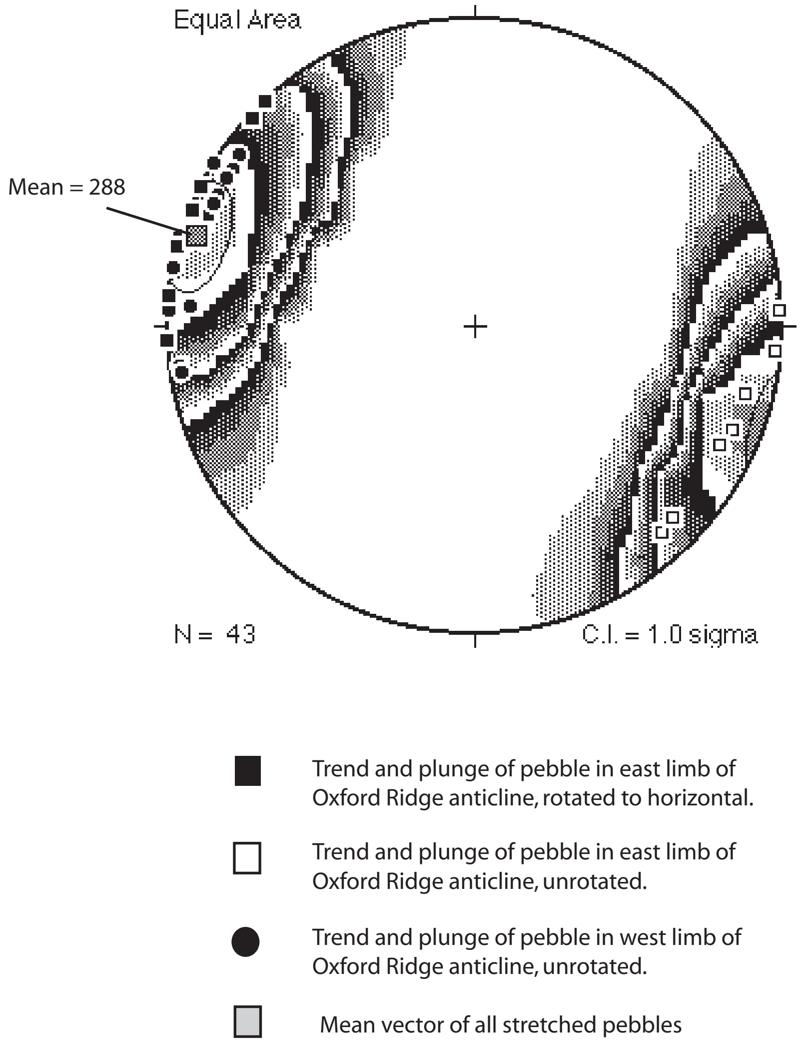


Figure 3-15. Stereogram showing trend and plunge of stretched pebbles in the Neoproterozoic Pocatello Formation. Kamb contours include all data sets.

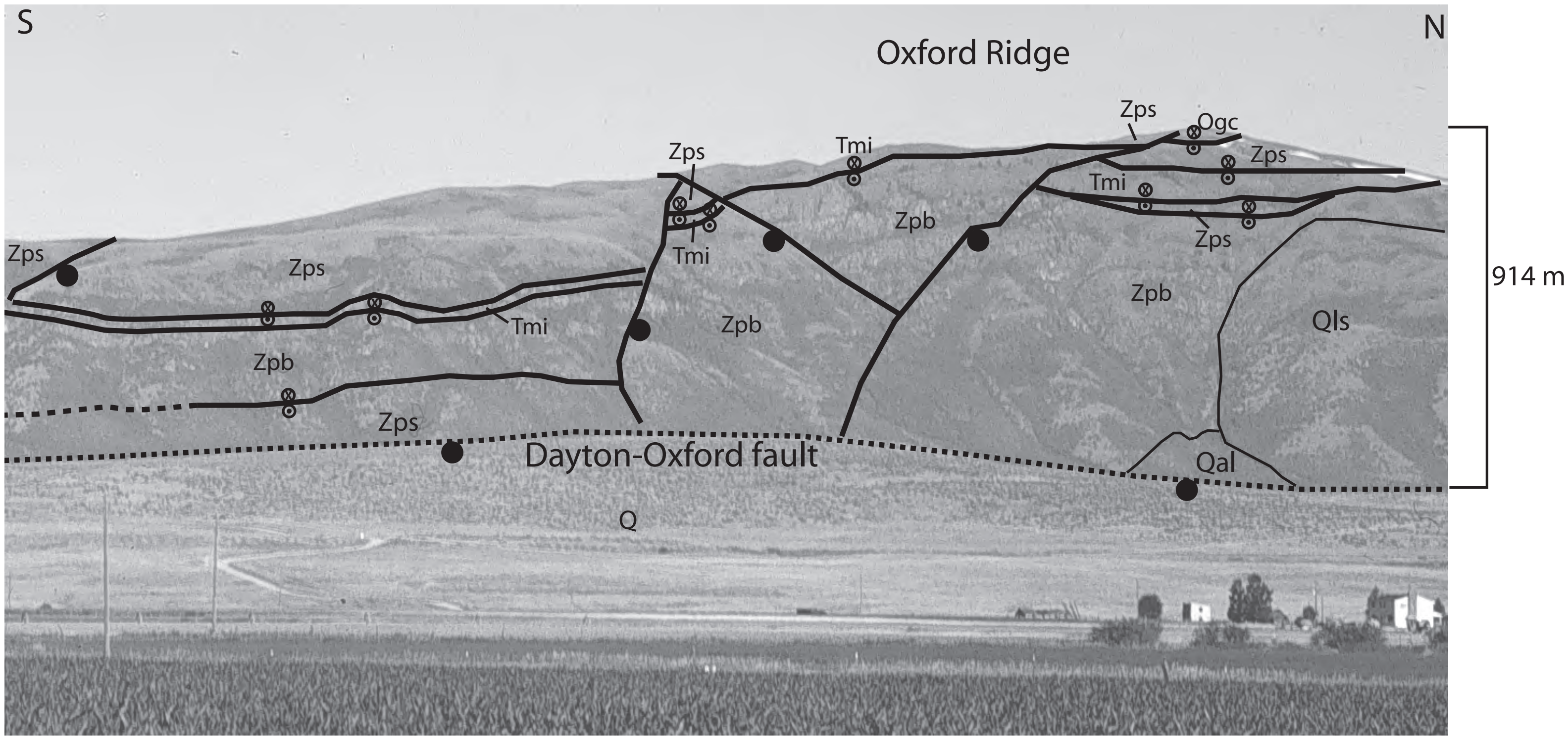


Figure 3-16. View west of low-angle normal faults on the east face of Oxford Ridge. These faults dip east and are cut by several younger normal faults. Abbreviations are: Ogc = Ordovician Garden City Formation, Tmi = Tertiary mafic intrusion, Zps = Pocatello Formation-Scout Mountain Member, Zpb = Pocatello Formation-Bannock Volcanic Member. (Height of ridge = 914 m)

Steep to moderately dipping normal fault with ball on downthrown side

Contact

Slip of low-angle fault away (into page)

Slip of low-angle fault toward (out of page)

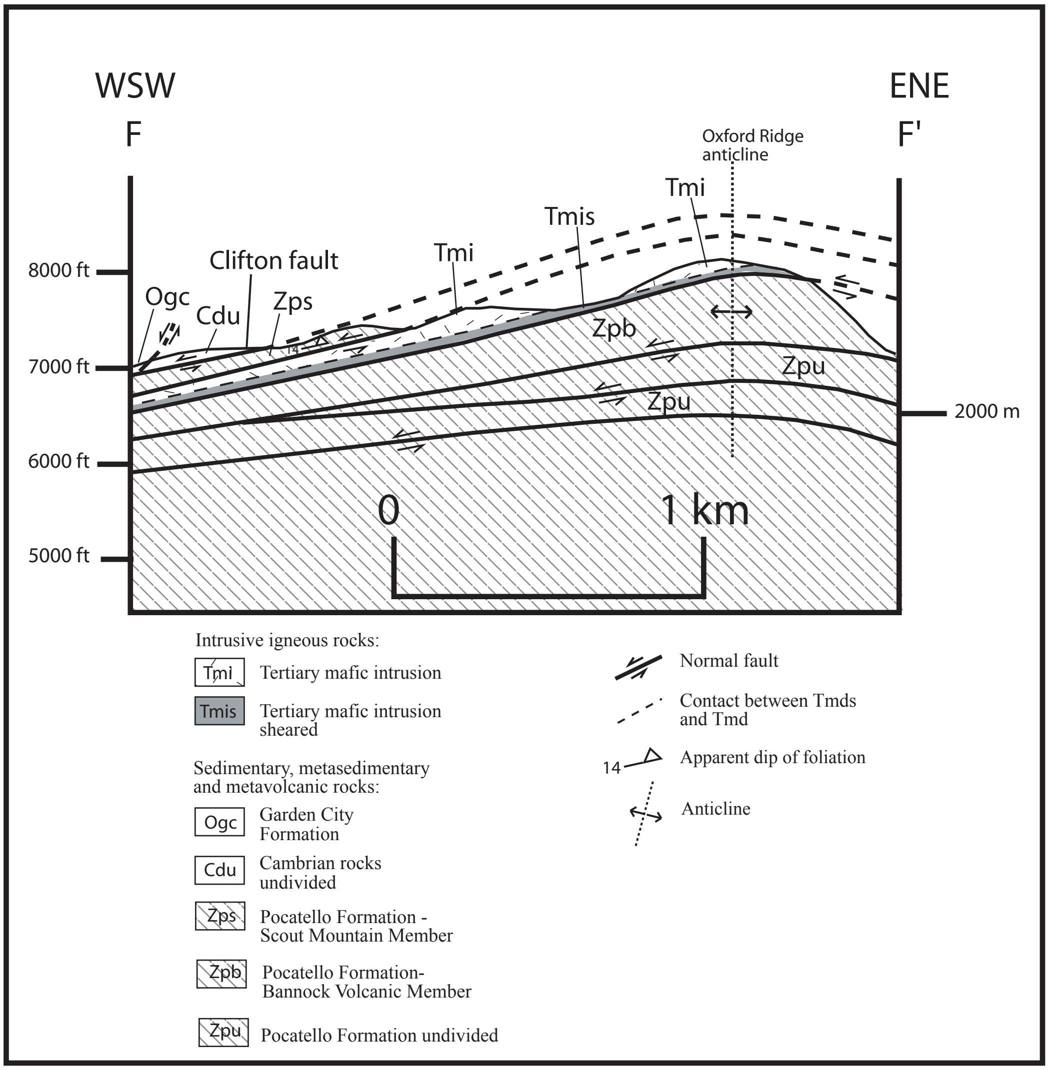


Figure 3-17. Geologic cross section F-F'. Section shows the Tertiary mafic intrusion intruded below the Clifton fault. See Figure 3 for location of section.

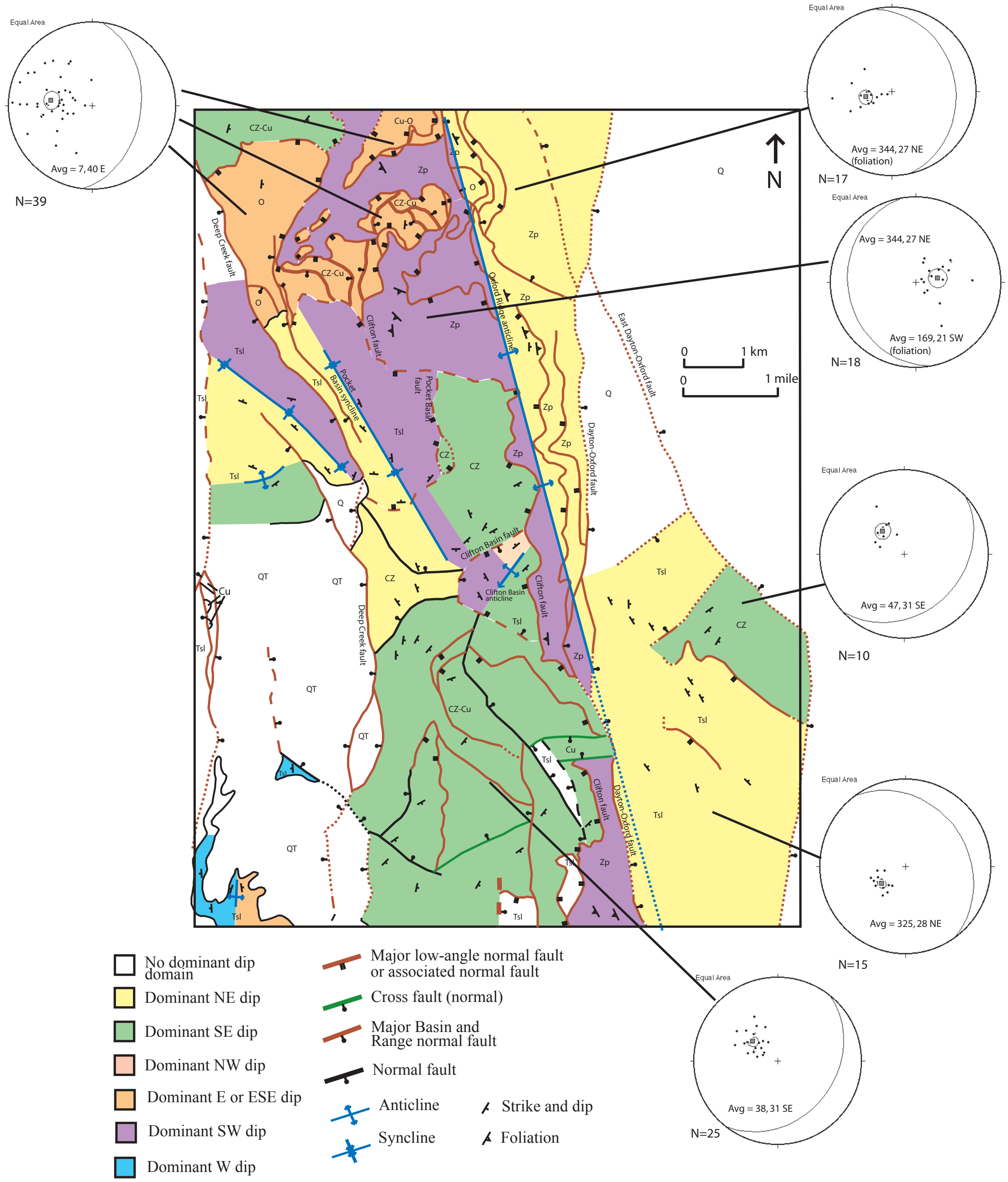


Figure 3-18. Dip-domain map with stereograms. Color-coded to show areas with dominant dip directions of strata in the Clifton quadrangle. Note that rocks in the footwall of the Clifton fault all strike NNW. Rocks in the hanging wall have a wide range of strikes. Abbreviations: Q = Quaternary, QT = Quaternary-Tertiary, Tsl = Tertiary Salt Lake Formation, O = Ordovician, Cu = Cambrian unddiferentiated, CZ-Cu = Cambrian and Brigham Group undifferentiated, CZ = Brigham Group, Zp = Neoproterozoic Pocatello Formation.

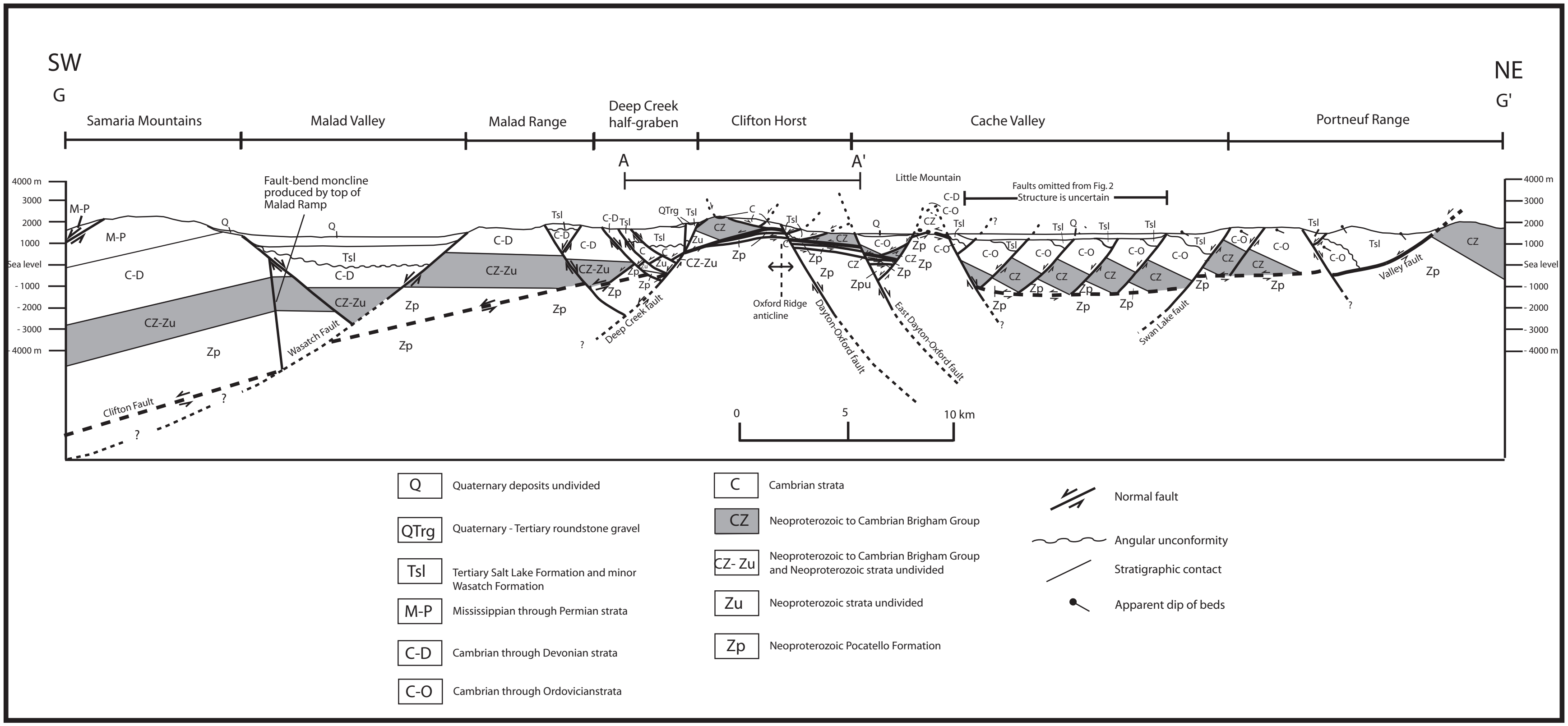


Figure 3-19a. Regional geologic cross section, G-G', of the northern Cache Valley, Idaho area. The structure beneath Cache Valley is uncertain. Mafic intrusions were omitted for simplicity. The Oxford Peak anticline is not evident on this line of section. Surface geology was compiled from Wach (1967), Platt (1977), Oriel and Platt (1968, 1980), Link and LeFebre (1983), Danzl (1985), Janecke and Evans (1999), Janecke et al. (in press), Janecke, unpublished mapping, and this study. Peterson and Oriel (1970) provided some subsurface control. Restoration indicated ~60% extension. See Figure 2 for location of this section and Figure 3 for locationof section A-A'.

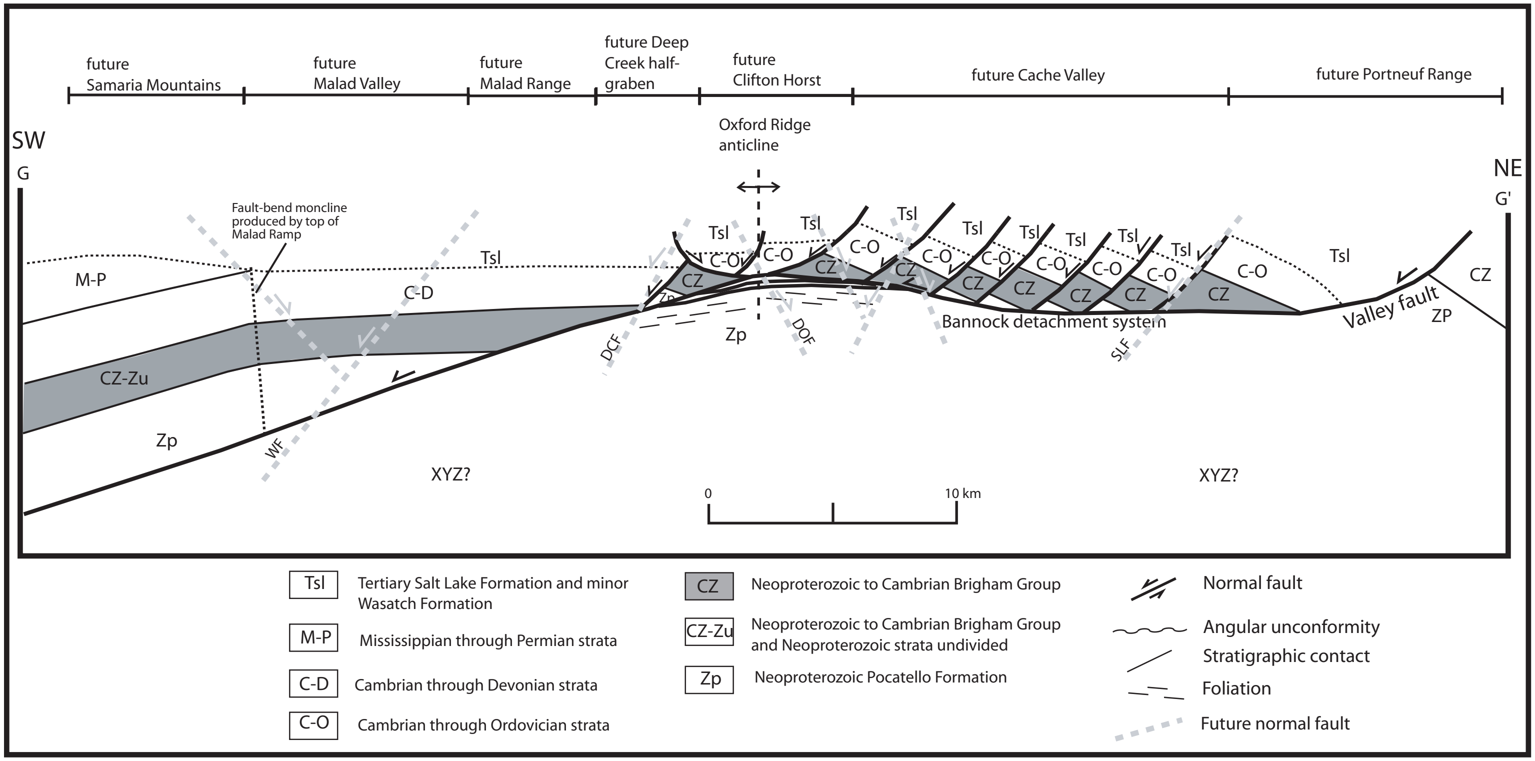


Figure 3-19b. Reconstruction of cross section G-G' shows Bannock detachment system before Basin-and-Range extension. Approximate future positions of Basin-and-Range topography and faults are shown. Reconstruction indicates that about 10% extension was accommodated by Basin-and-Range faults. Abbreviations: DCF = Deep Creek fault, DOF = Dayton-Oxford fault, SLF = Swan Lake fault, WF = Wasatch fault.

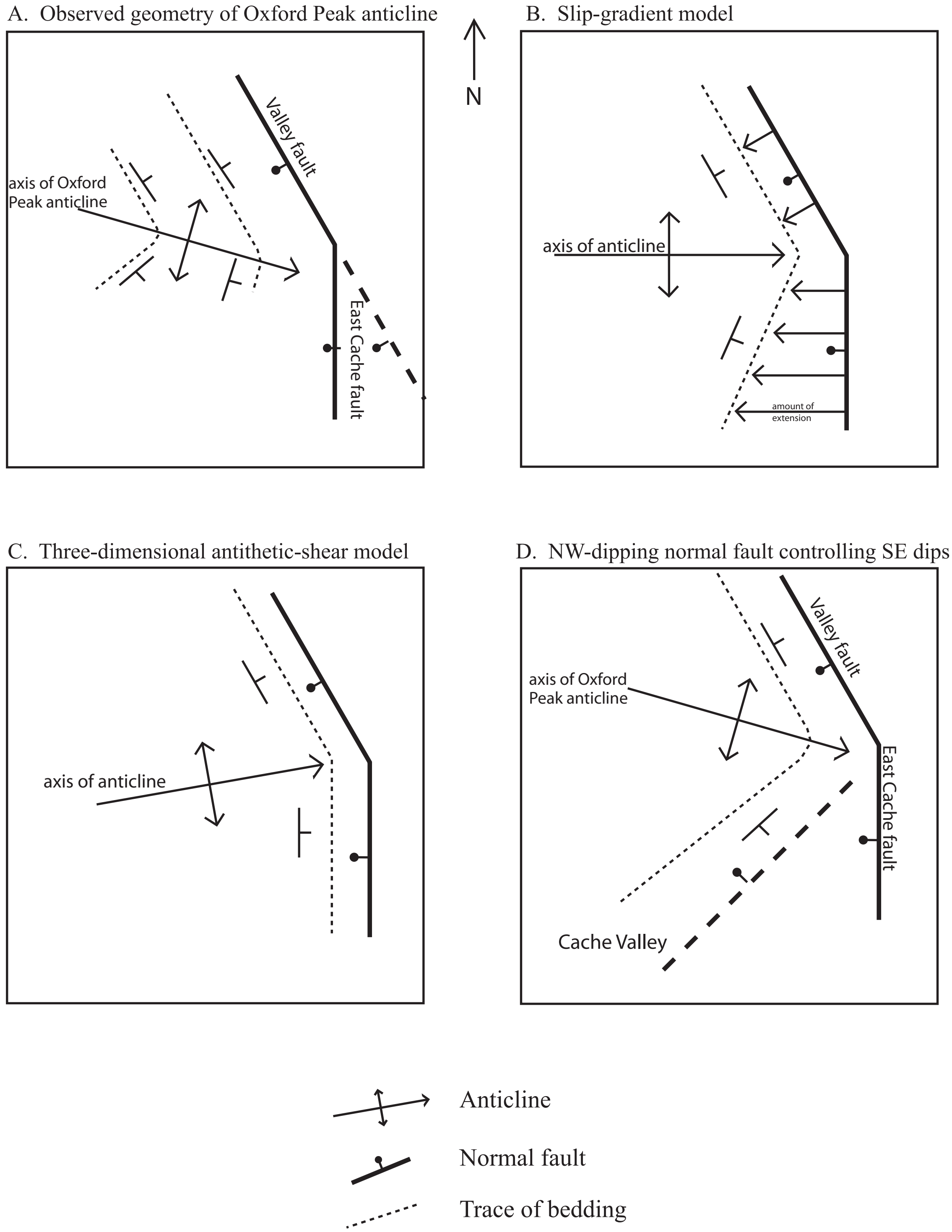


Figure 3-20. Schematic diagrams showing the observed geometry of the Oxford Peak anticline and three possible models for its origin.

CHAPTER 4

CONCLUSIONS

Two major episodes of extension are recorded in the southern Bannock Range of southeast Idaho. The first extensional episode was accommodated by the Bannock detachment system, a regional system that extends more than 130 km N-S. The Bannock detachment system collapsed the Cache-Pocatello culmination above and east of the Malad Ramp of the Paris-Willard thrust of the Sevier fold-and-thrust belt. Structural relationships mapped in the Clifton quadrangle indicate that the master detachment fault of the Bannock detachment system, the Clifton fault, is the youngest of the low-angle normal faults because it cut up, into its hanging wall, and excised part of the hanging wall. The Clifton fault was active at angles <25°, and did not rotate to lower angles through time. Compilation of regional map data indicates that the Valley fault in the Portneuf Range, east of the study area, is probably the breakaway for the detachment system exposed in the area of study.

Extension on the Bannock detachment system began before ~12 Ma with a cohesive hanging wall. The ESE-trending, Oxford Peak anticline was produced, possibly by three-dimensional antithetic shear, during initial WSW-directed translation of the Bannock detachment system. The Cache Valley Member of the Salt Lake Formation was deposited at this time in a large saline/alkaline lake in the hanging wall of the detachment system.

As extension on the Bannock detachment system continued, the hanging wall began to break up along high-angle, listric and planar normal faults, spaced as much as 2.8 km apart and as little as 200 m apart. These hanging-wall faults either were cut by or soled into the master detachment fault. This phase of extension produced many separate rift basins in the hanging wall of the detachment system, and thereby, transformed the depositional environment of the Salt Lake Formation. During this time, the Third Creek Member of the Salt Lake Formation was deposited in freshwater lakes and streams and deltas. Fault blocks in the hanging wall of the detachment system became emerging highlands and the source for clasts in conglomerates of the Third Creek Member. At some time during this phase of extension, a mafic sill invaded the footwall of the Bannock detachment system in the study area and the hanging wall of the detachment system farther south and northeast of the study area.

As the hanging wall continued to break up, depositional environments transitioned to braided streams. During this time, the footwall of the Bannock detachment system began to dome upward isostatically. Simultaneously, the master detachment fault, the Clifton fault, began to excise its hanging wall, and the Oxford Ridge anticline was formed in the footwall of the detachment system. The footwall of the detachment system is mostly Neoproterozoic Pocatello Formation that has been metamorphosed to greenschist and locally amphibolite facies. We interpret the Oxford Ridge anticline as the proto-core of the Bannock detachment system.

High-angle, E- and NE-striking normal faults cross-cut the Bannock detachment system. This brief period of faulting may be related to migration of the Yellowstone hotspot which was roughly north of the study area at this time, ~4-5 Ma. Following cross faulting, extension of the area was taken up on high-angle, Basin-and-Range normal faults. Initial uplift of the Clifton horst by the Dayton-Oxford and Deep Creek faults was followed by a period of quiescence during which time thin, Pliocene-Pleistocene(?) piedmont gravels were deposited. Renewed uplift of the Clifton horst resulted in the present topographic relief of the southern Bannock Range. The range-bounding faults cut and exposed the older Bannock detachment system and cross faults in the Clifton horst. The sedimentological signature of Basin-and-Range faulting (episode 3) is small.

Overall, 60% extension has occurred in the Cache Valley area of southeastern Idaho, about 50% accommodated by the Bannock detachment systems and about 10% accommodated by Basin-and-Range extension. This study shows that the Bannock detachment system has structural and stratigraphic characteristics of major low-angle normal-fault systems, such as the Sevier

Desert detachment system, in the Great Basin of the western United States. The system had evolved far enough to produce an incipient domal geometry and we interpret the Bannock detachment system to be a proto-metamorphic core complex of Miocene-Pliocene age. This study also shows an evolutionary sequence of structural and stratigraphic events commonly documented in the Basin-and-Range province: (1) Early, large-magnitude extension accommodated by a regionally extensive listric, low-angle normal-fault system; (2) Break-up of the hanging wall of the low-angle normal-fault system along closely spaced, listric and planar normal faults; (3) Small-magnitude extension accommodated by high-angle, Basin-and-Range normal faults that cut the earlier low-angle normal-fault system. Initial extension began in the study area in the middle to late Miocene, whereas, Basin-and-Range extension likely started after ~4-5 Ma, which is young for the Basin-and-Range province. Extension appears to migrate east and west in the Basin-and-Range province from its central axis (Perkins et al., 1998; Henry and Perkins, 2001).

References Cited

Henry, C.D., and Perkins, M.E., 2001, Sierra Nevada Basin and Range transition near Reno Nevada; two-stage development at 12 and 3 Ma: Geology, v. 29, p. 719-722.

Perkins, M. E., Brown, F. H., Nash, W. P., McIntosh, W., and Williams, S. K., 1998, Sequence, age, and source of silicic fallout tuffs in middle to late Miocene basins of the northern Basin and Range province: Geological Society of America Bulletin, v. 110, p. 344-360.

APPENDICES

Appendix A. Description of Map Units

Description of Map Units

Rock descriptions were collected while in the field.

Quaternary

Qafy Younger alluvial fan deposits - poorly sorted boulders, cobbles, pebbles, and sand that are post-Bonneville in age. Fans occur mostly on the east face of Oxford Ridge, along the Oxford-Dayton fault.

Qafo Older alluvial fan deposits - poorly sorted boulders, cobbles, pebbles, and sand that are pre-Bonneville in age. These fans are primarily located on the east face of Oxford Ridge and west of Weston Peak in the west-central area of the Clifton quadrangle.

Qal Alluvial deposits - fine-grained to gravelly stratified sediment, primarily deposited in stream channels.

Qbo Pleistocene near-shore sediments of the Bonneville level of the Bonneville lake cycle. The deposits consist of cobbles, pebbles, and sand.

Qbr Holocene flood-plain, and abandoned meanders of the ancestral Bear River. These deposits consist mostly of fine grained, river sediments.

Qc Colluvial deposits - very coarse to fine rock debris and soil material.

Qct Colluvial and talus deposits - very coarse to fine, angular rock debris and soil material.

Qls Landslide deposits - composition varies with parent material. These deposits may contain rotated blocks. Hachures indicate location of head scarp, where visible.

Qp Pleistocene near-shore sediments of the Provo level of the Bonneville lake cycle. The unit consists of mostly sand and mud, with cobbles and pebbles.

Qafp Pleistocene near-shore alluvial fan deposits from the subaerial portion of the fan-delta at the mouth of Clifton Creek. These deposits are graded to the Provo level of the Bonneville lake cycle. They consist mostly of cobbles and pebbles, but also contain some sand and mud.

QTg Quaternary-Tertiary (?) gravels - thin alluvial fan deposits (?) overlying Tertiary gravel (Tg) deposits and umconformably overlying the Third Creek Member of the Salt Lake Formation in the west and southwestern areas of the Clifton quadrangle. These gravels also overlie Paleozoic and Neoproterozoic bedrock within the Clifton horst. The gravel deposits reflect lithogys of nearby sources. In the Clifton Basin area the deposits are primarily composed of brecciated boulders and cobbles of Brigham Group quartzites. In the Weston Peak area the deposits are composed of brecciated and re-cemented Cambrian dolostones.

QTrg Quaternary-Tertiary roundstone gravel - Thin, gravel deposits exposed south of Second Creek in the southwestern area of the Clifton quadrangle. The deposits are in the hanging wall of the Deep Creek fault. The unit is composed of unlithified, uniformly well-rounded clasts consisting of mostly Brigham Group quartzites, Paleozoic carbonates and some recycled clasts from the Cache Valley member of the Salt Lake Formation.

Tertiary

Salt Lake Formation (Miocene - Pliocene)

Tnc **New Canyon member (late Miocene and early Pliocene(?) - 4.4 to 5.1 Ma)** Conglomerate exposed in the southeastern area of the Clifton quadrangle and north of Second Creek in the northwestern area of the Clifton quadrangle. Excellent consolidated exposures of this member northwest of the map area (in the Malad Summit quadrangle), west of Oxford Peak, show that the member is a parallel-bedded pebble to cobble conglomerate. The member is clast-supported and the clasts are uniformly well-rounded with no fining- or coarsening-upward sequences. Clasts consist of mostly Brigham Group quartzites, Paleozoic carbonates and some recycled clasts from the Cache Valley member of the Salt Lake Formation. The New Canyon member can be distinguished from the Third Creek member by its lack of tephras and freshwater limestones. Exact thickness of the New Canyon member is unknown in the map area. Reconnaissance mapping suggests that more than 1 km of section may be preserved west of Oxford Peak in the Malad Summit quadrangle to the north.

Ttc **Third Creek member (late Miocene - post-10.13 to pre-4.4 to 5.1 Ma)** Interbedded conglomerates, primary and slightly reworked tephras, and limestones. This member is exposed on the eastern side of the Clifton horst in the hanging wall of the Oxford-Dayton fault and on the western side of the Clifton horst in the hanging wall of the Deep Creek fault. It is also exposed within the Clifton horst in the south-central section of the Clifton quadrangle. It is composed mostly of poorly consolidated white to gray tephras, slightly tuffaceous and locally oolitic and fossiliferous limestones, medium to course-grained sandstones, and clast- and matrix-supported conglomerate. Matrix of the conglomerate is variably tuffaceous, sandy, and calcareous. Individual conglomerate beds range from < 50 cm to > 4 meters in thickness. Tephra and limestone beds vary from < 20 cm to > 2 meters. Clasts within conglomerate beds are generally subangular- to well-rounded pebbles, but a few beds contain cobbles. Clasts consist primarily of Brigham Group quartzites, Paleozoic carbonates and clasts recycled from the Cache Valley Member of the Salt Lake Formation. Minimum thickness of this member in the map area is 610 m.

Tcv **Cache Valley member (late Miocene - pre-10.3 to post-10.13)** The majority of this unit is composed of slightly reworked tuffaceous sedimentary rocks, including mudstone, siltstone, limestone, silicified limestone, shale, sandstone, and rare pebbly conglomerates, as well as some primary tuffs and tephras. The Cache Valley member is primarily exposed on west of Oxford-Ridge in the hanging wall of the Clifton and New Canyon detachment faults. Color is variable, but light brown, off-whites, and light green rocks are most common. The rocks are generally characterized by a silica or clay cement and are more indurated than the tuffaceous rocks of the overlying Third Creek member of the Salt Lake Formation. Minimum thickness is 600 m in the map area.

Tmi **Tertiary Mafic Intrusion (Upper Tertiary?)** Dark green, coarse-grained, metamorphosed diabase(?). According to Raymond (1971), the mafic intrusion is amphibolite with 50 percent hornblend and lesser amounts of epidote, chlorite, leucoxene, sericite, hematite, and clay. Our analysis showed abundant amounts of plagioclase. The pluton is located on Oxford Ridge in the northern area of the Clifton quadrangle. It forms a sill-like WSW-dipping tabular body that is parallel to Cenozoic foliation in the Neoproterozoic Pocatello Formation and appears to intrude and follow Tertiary low-angle normal faults. The upper-most part of the sill contains brittle fault rocks that are intruded by less deformed felsic restite of the mafic sill. This deposit has not been dated. It grossly resembles Miocene-Pliocene diabase dikes and sills that intruded the Tertiary Salt Lake Formation in the northeast and west-central Cache Valley (Willard, 1972; Winter, 1985; Biek et al., 2001).

Tmis **Tertiary Mafic Intrusion – Sheared (Upper Tertiary?)** Silvery-gray, medium- to fine-grained, sheared, metamorphosed mafic intrusion. Found in the deepest exposures of the pluton located on Oxford Ridge in the northern area of the Clifton quadrangle.

Tw **Wasatch Formation (Paleocene and Eocene)** A red sandy to conglomeratic deposit composed primarily of Paleozoic carbonate clasts. It is exposed in isolated fault-blocks west of Oxford Ridge and generally overlies the Ordovician Garden City Formation in unconformtiy. The red matrix of this deposit easily distinguishes it from the overlying Cache Valley member of the Salt Lake Formation. The minimum thickness of the Wasatch Formation in the map area is 13 m, but it is as thick as 35m in the Malad City East quadrangle to the west and may be as thick as 200 m to the southwest in the Henderson Creek quadrangle.

Silurian

SOf **Fish Haven Dolomite (Lower Silurian and Upper Ordovician)** Dark gray to black dolomite with thin, light gray chert beds and some chert nodules. A small fault block of Fish Haven is exposed on Oxford Ridge in the Clifton quadrangle.

Ordovician

Ogc **Garden City Limestone (Middle and Lower Ordovician)** Gray, thin- to medium-bedded, fossiliferous limestone containing many intraformational conglomerate beds that sometimes weather to a reddish color; black chert nodules abundant near the top. A complete section not exposed; minimum thickness estimated at 55 m.

Cambrian

Cdu **Cambrian Dolostone undifferentiated (Upper Cambrian)** Dark gray to light gray dolostone exposed in fault blocks on and west of Oxford Ridge in the Clifton quadrangle. These dolostones may be either the Cambrian Nounan Formation or Cambrian St. Charles Formation and very locally may include the Silurian-Ordovician Fish Haven Formation.

Csc **St. Charles Formation (Upper Cambrian)** The lower section consists of medium to thin-bedded, medium to dark gray limestone with interbeds of silicified siltstone and chert. Fossil hash is found throughout. The upper section consists of thick-bedded, coarsely crystalline, light to medium gray dolostone with local white to tan colored chert nodules and stringers. Complete section not exposed; minimum thickness estimated at 90 m.

Cwc **Worm Creek Quartzite Member of the St. Charles Formation (Upper Cambrian)** Gray to tan to pink arkosic sandstone with minor interbedded dolomite and limestone; quartzite beds have distinctive chalky-weathering feldspar grains. Minimum thickness is 19 m.

Cnu **upper Nounan Formation (Upper and Middle Cambrian)** Thin to medium bedded, medium to dark gray, silty limestone and light-gray, sugary dolomite with interbedded tan to red siltstone and fine-grained sandstone. A complete section is not exposed, but minimum thickness 490 m.

Cnl **lower Nounan Formation (Upper and Middle Cambrian)** Cliff-forming thick-bedded, coarsely crystalline, light to medium-gray, fenestral dolostone.

Cn **Nounan Formation, undivided (Upper and Middle Cambrian)** Mostly cliff-forming and ledge-forming, thin to thick bedded, dark to medium gray dolostone and medium to dark gray limestone with sparsely interbedded tan siltstone.

Cbo **Bloomington Limestone (Middle Cambrian)** The top of the unit consists of brown-weathering, slope-forming, light green and tan shale with interbedded , thin bedded, medium-gray limestone and intraformational limestone conglomerate. These shales contain distinctive green-gray, micritic, limestone nodules. The middle of the unit consists of ledge-forming, medium-bedded, medium to light-gray, locally oolitic limestone with interbedded, intraformational limestone conglomerate. The lower unit consists of brownish-orange weathering, slope forming, brownish-gray shale with interbedded, thin-bedded, oolitic limestone. A complete section is not exposed, but minimum thickness is 460 m.

Cbl **Blacksmith Formation (Middle Cambrian)** Dark-gray, hackly weathering, cliff- to ledge-forming, limestone with minor interbedded, light-gray dolostone; contains light-gray to pale-orange silty bands and mottling, and is locally oolitic with few oncolites. A complete section is not exposed, but minimum thickness east of Weston Peak is 380 m.

Clb **Lead Bell Shale (Middle Cambrian)** Slope-forming, light-gray, silty, oncolite-bearing limestone and tan, silty dolostone with interbedded tan and red fine-grained, micaceous siltstone and thin-bedded, silvery shale. This unit is equivalent to the Langston and Ute Formations of Utah stratigraphy. Thickness in the map area is unknown.

Cwp **Windy Pass Argillite (Lower Cambrian)** Dark gray to brown, micaceous mudstone and siltstone, and dark brown weathering, gray quartzite. A complete section is not exposed, but minimum thickness in the south-central area of the Clifton quadrangle is estimated to be 300 m.

CZcm **Camelback Mountain Quartzite (Lower Cambrian and Upper Proterozoic)** Upper Camelback Mountain Quartzite Formation - light pink to white color with green partings, fine to coarse grained, vitreous sandstone. Lower Camelback Mountain Quartzite Formation - Dark pink to maroon color, liesegang banded, coarse-grained to pebbly sandstone and conglomerate at the base. Minimum thickness estimated to be 430 m.

Proterozoic

Zrp **Rocky Peak Phyllite Member of the Camelback Mountain Quartzite Formation (Upper Proterozoic)** Dark purple, fine-grained, micaceous siltstone with well developed cleavage and reduction spots. The Rocky Peak Phyllite Member is exposed in the Davis Basin area and on the southwest side of the Clifton horst near Old Baldy Peak in the Clifton quadrangle.

Zm **Mutual Formation (Upper Proterozoic)** Maroon to pinkish purple, medium to coarse grained quartzite; locally conglomeratic with rare, dark purple, argillite interbeds. The Mutual Formation is exposed in the Davis Basin area and on the southwest side of the Clifton horst near Old Baldy Peak in the Clifton quadrangle. It is also exposed in a small fault block on Oxford Ridge. The base of the Mutual Formation is not exposed and thickness in map area unknown.

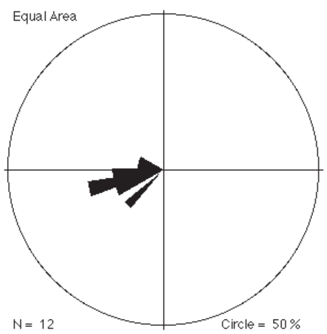
Zps **Pocatello Formation - Scout Mountain Member (Neoproterozoic)** Slope and ledge forming, thick to massive bedded, brown and green diamictite, brown siltstone and sandstone and conglomerate metamorphosed to greenschists facies. The diamictite is a pebble to cobble-bearing, matrix supported mudstone to fine sandstone. Clasts of the diamictite are locally stretched. The Scout Mountain Member is exposed on the east side of the Clifton horst and on and west of Oxford Ridge.

Zpb **Pocatello Formation - Bannock Volcanic Member (Neoproterozoic)** Ledge and cliff-forming, massive metavolcanic breccia and metabasalt. Metabasalt is a greenstone composed of chlorite, quartz, and epidote and contains pillow lavas and locally preserved vesicles.

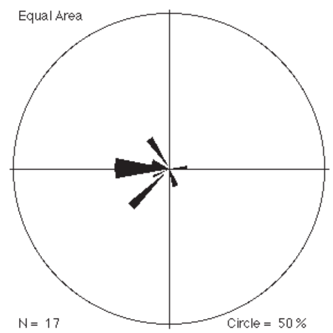
Zps? **Pocatello Formation - sheared Scout Mountain Member ? (Neoproterozoic)** Very sheared, phyllitic, light green weathering outcrops. The parallel foliation is so pervasive that the protolith is uncertain; either sedimentary or volcanic rocks may be the protolith. The sheared Scout Mountain Member(?) is topographically and stratigraphically lower than the Bannock Volcanic Member and is exposed on the east-facing slope of Oxford Ridge.

Appendix B. Paleocurrent Data from the Third Creek Member of the Salt Lake Formation

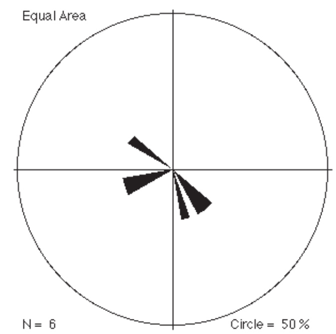
Rose diagrams showing paleocurrent direction from imbricated pebbles and fluvial troughs and scours in the Third Creek Member of the Salt Lake Formation. Site number location is indicated on the geologic map (Plate 1). A complete list of data is provided in the table following.



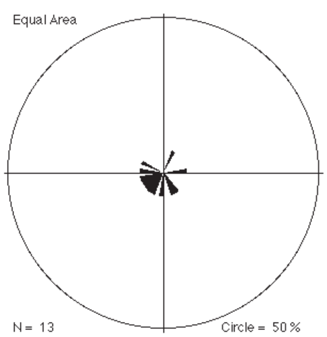
Site 402 - imbrication
East side of Clifton horst
Near top of Third Creek Member



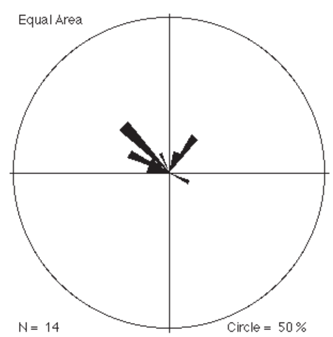
Site 403 - imbrication
East side of Clifton horst
Near top of Third Ceek Member



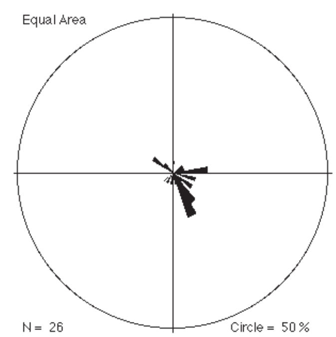
Scour axes
East side of Clifton horst
Near top of Third Creek Member



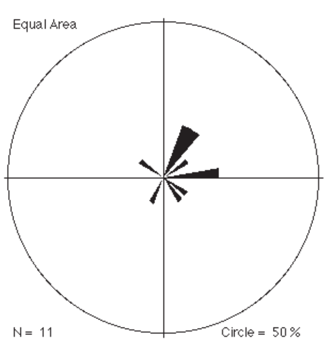
Site 407a - imbrication
West side of Clifton horst
Middle Third Creek Member



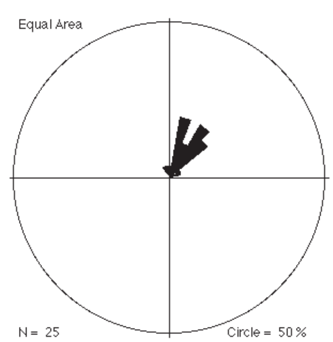
Site 407b - imbrication
West side of Clifton horst
Middle Third Creek Member



Site 407c - imbrication
West side of Clifton horst
Middle Third Creek Member



Site 407d - imbrication
West side of Clifton horst
MIddle Third Creek Member



Site 407e - imbrication
West side of Clifton horst
Middle Third Creek Member

Paleocurrent data from the Third Creek Member of the Salt Lake Formation

Site numbers are indicated on the geologic map (Plate 1). All data listed follow right-hand rule.

Scour Data

Site number	Location	Strike/dip bedding	Data: Strike/dip of Limb; Trend/plunge of rotated axis
400	East side of horst, Near top of Ttc	313, 30 E	Limbs: 10, 31 E; 328, 27 E T/P of rotated axis: 244, 1 SW
403a	East side of horst, Near top of Ttc	322, 21 E	Limbs: 261, 30 N; 76, 40 E T/P of rotated axis: 259, 16 W
403b	East side of horst, Near top of Ttc	322, 21 E	Limbs: 50, 39 E; 76, 40 SE T/P of rotated axis: 144, 41 S
403c	East side of horst, Near top of Ttc	322, 21 E	Limbs: 337, 38 E; 312, 12 E T/P of rotated axis: 166, 2 SW
403d	East side of horst, Near top of Ttc	322, 21 E	Limbs: 342, 29 E; 87, 14 S T/P of rotated axis: 137, 10 S
406	East side of horst, Near top of Ttc	334, 32 E	Measured axis: 287, 18 N T/P of rotated axis: 300, 39 NW

Imbricate Data

Site number	Location	Strike/dip bedding	Strike/dip pebble	Trend/Plunge of pole to plane of pebble corrected for attitude of bedding
402	East side of horst, Near top of Ttc	304, 11 E	341, 31	266, 67
			339, 52	255, 47
			004, 46	285, 49
			317, 20	242, 80
			324, 39	241, 61
			335, 40	254, 59
			330, 26	256, 73
			355, 87	266, 10
			009, 34	296, 59
			313, 26	229, 75
			352, 33	278, 63
			312, 34	255, 67
403	East side of horst, Near top of Ttc	322, 21 E	355, 41	288, 64
			333, 24	292, 85
			205, 23	87, 53
			312, 22	153, 86
			297, 27	161, 78
			029, 41	325, 53
			337, 34	267, 75
			326, 30	245, 81
			317, 80	226, 31
			355, 23	328, 78
			350, 45	277, 62
			319, 48	227, 63
			229, 40	264, 69
			345, 46	269, 62
			350, 55	271, 53
			352, 46	279, 61
			318, 48	226, 63

407a	West side of horst, Middle Ttc	290, 10 NE	292, 41 280, 28 246, 34 254, 40 318, 22 304, 16 001, 20 194, 21 345, 24 306, 51 115, 34 335, 35 306, 36	203, 59 185, 72 143, 62 155, 58 247, 76 235, 83 299, 71 081, 66 278, 70 219, 49 024, 46 258, 61 221, 64
407b	West side of horst, Middle Ttc	264, 15 W	347, 21 259, 16 330, 19 011, 20 140, 28 005, 16 005, 27 055, 22 175, 13 125, 16 125, 23 341, 30 016, 20 356, 15	293, 66 118, 88 285, 71 310, 62 033, 52 312, 66 299, 57 336, 54 036, 70 016, 61 020, 54 277, 60 313, 61 309, 69
407c	West side of horst, Middle Ttc	303, 14 NE	246, 51 298, 43 235, 22 269, 35 245, 43 234, 31 165, 9 015, 38 089, 45 025, 60 271, 26 265, 30 266, 18 273, 26 210, 16 010, 22 219, 31 260, 31 180, 25 220, 15 199, 23 175, 20 310, 25 280, 51 250, 31 265, 34	145, 46 206, 61 109, 69 162, 66 140, 53 120, 61 050, 69 304, 54 005, 33 303, 31 155, 74 152, 69 126, 79 158, 75 081, 68 316, 69 106, 58 147, 67 071, 56 085, 71 083, 61 065, 59 228, 79 184, 52 136, 65 156, 66

407d	West side of horst, Middle Ttc	300, 62 E	306, 38 283, 38 299, 68 295, 39 300, 60 270, 49 274, 65 290, 66 270, 48 301, 62 306, 38	021, 66 053, 63 201, 84 038, 67 30, 88 84, 62 121, 67 142, 80 082, 62 300, 89 021, 66
407e	West side of horst, Middle Ttc	282, 35 NE	270, 10 240, 15 263, 6 229, 18 154, 7 229, 17 312, 26 203, 42 192, 41 215, 26 195, 10 324, 15 304, 15 124, 22 102, 16 303, 25 286, 8 301, 6 282, 30 260, 19 210, 11 235, 17 249, 17 150, 27 252, 22	017, 65 36, 64 016, 61 044, 62 021, 50 42, 62 325, 73 076, 43 069, 38 059, 57 029, 54 348, 64 357, 68 022, 34 012, 39 334, 76 011, 63 008, 61 012, 85 034, 71 031, 57 041, 64 037, 68 036, 34 047, 71

Appendix C. Release Letters

November 4, 2002

Stephanie Carney
Utah State University
Logan, UT 84321-4505
(435) 797-1273

Jeff Evans
N7653 Island Lake Rd.
Spooner, WI 54801

Dear Mr. Evans:

I am in the process of preparing my Master's thesis in the Department of Geology at Utah State University. I hope to complete my degree in the Fall of 2002.

I am requesting your permission to include in my thesis a journal article of which you are a co-author. The title of the article is *Late Miocene-Pliocene Detachment Faulting and Pliocene-Recent Basin-and-Range Extension Inferred from Dismembered Rift Basins of the Salt Lake Formation, Southeast Idaho*. This paper was submitted in May 2002 for publication in a special publication of the Society for Sedimentary Geology (SEPM). You will be cited as a co-author in the thesis chapter in which the paper will appear and this permission letter will be included in an appendix of the thesis.

Please indicate your approval of this request by signing in the space provided, then fax the signed form to 435-797-1588. If you have any questions, please call me at the number above.

Thank you,

Stephanie Carney

As a co-author of the following article, I hereby give permission to Stephanie Carney to use the article in her Master's thesis.

Janecke, S.U., Carney, S.M., Perkins, M.E., Evans, J.C., Link, P.K., Oaks, R.Q., Jr., and Nash, B.P., in press, Late Miocene-Pliocene detachment faulting and Pliocene-Pleistocene Basin-and-Range extension inferred from dismembered rift basins of the Salt Lake Formation, southeast Idaho, *in* Raynolds, R., and Flores, R., eds., Cenozoic Paleogeography of the Rocky Mountains: SEPM Special Publication.

Signed Jeff C. Evans

November 4, 2002

Stephanie Carney
Utah State University
Logan, UT 84321-4505
(435) 797-1273

Barbara P. Nash, Professor
Department of Geology and Geophysics
University of Utah
Salt Lake City, UT 84112-0011
(435) 581-8587

Dear Dr. Nash:

I am in the process of preparing my Master's thesis in the Department of Geology at Utah State University. I hope to complete my degree in the Fall of 2002.

I am requesting your permission to include in my thesis a journal article of which you are a co-author. The title of the article is *Late Miocene-Pliocene Detachment Faulting and Pliocene-Recent Basin-and-Range Extension Inferred from Dismembered Rift Basins of the Salt Lake Formation, Southeast Idaho.* This paper was submitted in May 2002 for publication in a special publication of the Society for Sedimentary Geology (SEPM). You will be cited as a co-author in the thesis chapter in which the paper will appear and this permission letter will be included in an appendix of the thesis.

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Signed Barbara Nash 11/4/02

November 4, 2002

Stephanie Carney
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Michael E. Perkins, Research Assistant Professor
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Salt Lake City, UT 84112-0011
(435)-581-6552

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Signed Michael E. Perkins 5 Nov 2002