BRIGHAM YOUNG UNIVERSITY

GEOLOGY

S T U D I E S

GEOLOGICAL SOCIETY OF AMERICA

FIELD TRIP GUIDE BOOK

1997 ANNUAL MEETING • SALT LAKE CITY, UTAH

PART 1 ONE

EDITED BY PAUL KARL LINK AND BART J. KOWALLIS

VOLUME 42 • 1997

PROTEROZOIC TO RECENT STRATIGRAPHY, TECTONICS, AND VOLCANOLOGY, UTAH, NEVADA, SOUTHERN IDAHO AND CENTRAL MEXICO

Edited by

Paul Karl Link and Bart J. Kowallis

BRIGHAM YOUNG UNIVERSITY GEOLOGY STUDIES

Volume 42, Part I, 1997

CONTENTS

A Publication of the
Department of Geology
Brigham Young University
Provo, Utah 84602

Editor

Bart J. Kowallis

Brigham Young University Geology Studies is published by the Department of Geology. This publication consists of graduate student and faculty research within the department as well as papers submitted by outside contributors. Each article submitted is externally reviewed by at least two qualified persons.

Cover photos taken by Paul Karl Link.

Top: Upheaval Dome, southeastern Utah.

Middle: Lake Bonneville shorelines west of Brigham City, Utah.

Bottom: Bryce Canyon National Park, Utah.

ISSN 0068-1016
9-97 700 23348/24218

Preface

Guidebooks have been part of the exploration of the American West since Oregon Trail days. Geologic guidebooks with maps and photographs are an especially graphic tool for school teachers, University classes, and visiting geologists to become familiar with the territory, the geologic issues and the available references.

It was in this spirit that we set out to compile this two-volume set of field trip descriptions for the Annual Meeting of the Geological Society of America in Salt Lake City in October 1997. We were seeking to produce a quality product, with fully peer-reviewed papers, and user-friendly field trip logs. We found we were bucking a tide in our profession which de-emphasizes guidebooks and paper products. If this tide continues we wish to be on record as producing "The Last Best Geologic Guidebook."

We thank all the authors who met our strict deadlines and contributed this outstanding set of papers. We hope this work will stand for years to come as a lasting introduction to the complex geology of the Colorado Plateau, Basin and Range, Wasatch Front, and Snake River Plain in the vicinity of Salt Lake City. Index maps to the field trips contained in each volume are on the back covers.

Part 1 "Proterozoic to Recent Stratigraphy, Tectonics and Volcanology: Utah, Nevada, Southern Idaho and Central Mexico" contains a number of papers of exceptional interest for their geologic synthesis. Part 2 "Mesozoic to Recent Geology of Utah" concentrates on the Colorado Plateau and the Wasatch Front.

Paul Link read all the papers and coordinated the review process. Bart Kowallis copy edited the manuscripts and coordinated the publication via Brigham Young University Geology Studies. We would like to thank all the reviewers, who were generally prompt and helpful in meeting our tight schedule. These included: Lee Allison, Genevieve Atwood, Gary Axen, Jim Beget, Myron Best, David Bice, Phyllis Camilleri, Marjorie Chan, Nick Christie-Blick, Gary Christenson, Dan Chure, Mary Droser, Ernie Duebendorfer, Tony Ekdale, Todd Ehlers, Ben Everitt, Geoff Freethey, Hugh Hurlow, Jim Garrison, Denny Geist, Jeff Geslin, Ron Greeley, Gus Gustason, Bill Hackett, Kimm Harty, Grant Heiken, Lehi Hintze, Peter Huntoon, Peter Isaacson, Jeff Keaton, Keith Ketner, Guy King, Mel Kuntz, Tim Lawton, Spencer Lucas, Lon McCarley, Meghan Miller, Gautam Mitra, Kathy Nichols, Robert Q. Oaks, Susan Olig, Jack Oviatt, Bill Perry, Andy Pulham, Dick Robison, Rube Ross, Rich Schweickert, Peter Sheehan, Norm Silberling, Dick Smith, Barry Solomon, K.O. Stanley, Kevin Stewart, Wanda Taylor, Glenn Thackray and Adolph Yonkee. In addition, we wish to thank all the dedicated workers at Brigham Young University Print Services and in the Department of Geology who contributed many long hours of work to these volumes.

Paul Karl Link and Bart J. Kowallis, Editors

Neoproterozoic Sedimentation and Tectonics in West-central Utah

NICHOLAS CHRISTIE-BLICK
Department of Earth and Environmental Sciences and Lamont-Doherty Earth Observatory of Columbia University, Palisades, New York 10964-8000

ABSTRACT

The principal themes of the excursion are clastic sedimentology and sequence stratigraphy in glacial-marine and fluvial to shallow marine settings; and stratigraphic evidence for crustal extension and passive-margin development in the western United States during Proterozoic and early Paleozoic time. The main focus of this guide is geology of Neoproterozoic age (1,000–543 Ma) in the Sheeprock Mountains and central Wasatch Range in west-central Utah. The third day is co-ordinated with a one-day excursion to the Big Cottonwood Formation (Ehlers et al., 1997), which may include rocks as old as Mesoproterozoic.

Highlights of the excursion are: 1) an expanded section of Sturtian-age (~750 Ma), deep-water glacial-marine conglomerate and diamictite, including dropstones and at least one striated clast (Sheeprock Group, Day 1); 2) glacially eroded valleys as much as 900 m deep, partially filled by the broadly correlative Mineral Fork Formation, and in one place with glacial grooves preserved at the contact (Day 3); and 3) incised valleys several tens to 160 m deep at three stratigraphic levels in the Brigham Group, these thought to be related to Varanger-age (~600 Ma) glacial-eustatic drawdown, and filled by a combination of coarse fluvial conglomerate, pebbly quartzite and marine siltstone (Days 2 and 3). Stratigraphic evidence for rifting consists of abrupt thickening of stratigraphic units towards the presumed locations of basin-bounding faults (Sheeprock Group), abrupt thinning across inferred faults (Big Cottonwood Formation), and truncation of tilted strata beneath the Tintic Quartzite (Lower Cambrian) in the central Wasatch Range.

OVERVIEW OF EXCURSION

Thick successions of Neoproterozoic and early Paleozoic age are exposed discontinuously throughout the North American Cordillera, from eastern Alaska to Sonora, Mexico (Crittenden et al., 1971; Stewart, 1972, 1982; Stewart and Poole, 1974; Burchfiel and Davis, 1975; Stewart and Suczek, 1977; Young, 1982; Christie-Blick and Levy, 1989a; Levy and Christie-Blick, 1991a; Link et al., 1993; Rainbird et al., 1996). The rocks record a protracted history of continental extension and passive-margin development, and they accumulated in a range of basinal to ramp and platformal settings. Sedimentary rocks of glacial and glacial-marine origin are present at two levels in the Neoproterozoic (~750 Ma and ~600 Ma), and are thought to correspond with the Sturtian and Varanger events known widely from many continents (Aitken, 1991; Hein and McMechan, 1994; Link et al., 1994; Ross et al., 1995).

Elements of this geology will be examined in the course of this excursion at two locations in west-central Utah: the Sheeprock Mountains, approximately 100 km south-southwest of Salt Lake City, and the central Wasatch Range, about 20 km southeast of the city (Fig. 1). The rocks were disrupted by folding and thrusting in the late Cretaceous (Armstrong, 1968; Crittenden, 1976; Christie-Blick, 1983a; Morris, 1983; Bruhn et al., 1986; Levy and Christie-Blick, 1989; Allmendinger, 1992) and by extensional block-faulting during the late Cenozoic (Stewart, 1978; Eaton, 1982; Smith and Bruhn, 1984; Wernicke, 1992). However, in many places sedimentary features are well preserved in rocks of greenschist metamorphic grade.

The localities to be visited are notable in several respects. The Neoproterozoic and Lower Cambrian succession in the Sheeprock Mountains is amongst the thickest (> 7 km) and most complete in the western United States, including more than 2 km of Sturtian-age glacial-marine strata at the base (Fig. 2; Christie-Blick, 1982, 1983a; Crittenden et al., 1983; Link et al., 1994). Correlative strata in the central Wasatch Range are comparatively thin (< 1,400 m), but glacial deposits rest on a surface with as much as 900 m of erosional relief (Crittenden et al., 1952; Crittenden, 1976; Ojakangas and Matsch, 1980; Christie-Blick, 1983b;

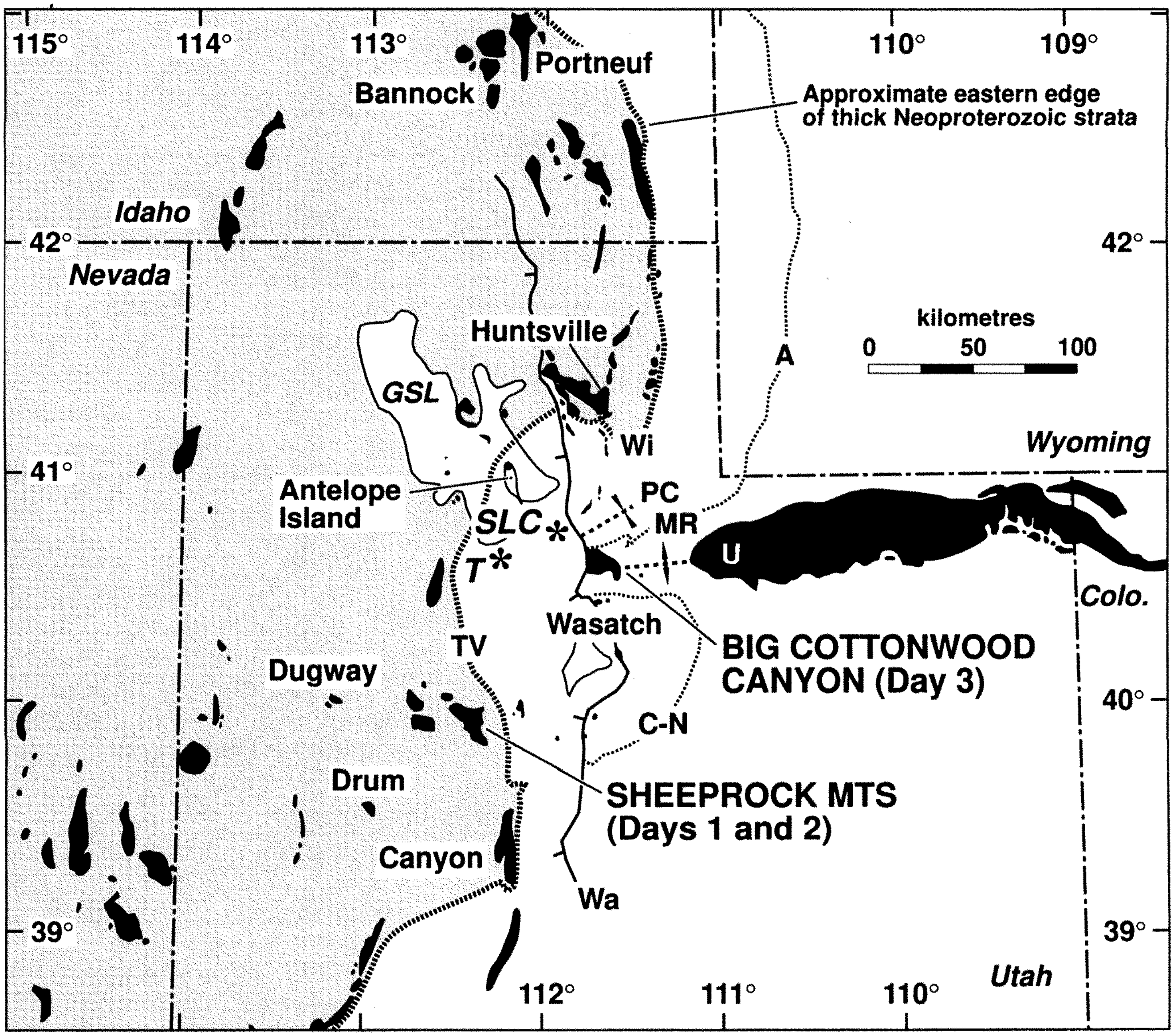

Figure 1. Tectonic setting and distribution of Mesoproterozoic to Lower Cambrian outcrops (shown in black) in west-central Utah and adjacent areas, and location of field excursion traverses and stops in the Sheeprock Mountains and central Wasatch Range (Big Cottonwood Canyon). Selected thrust faults: A, Absaroka; MR, Mount Raymond; Wi, Willard; C-N, Charleston-Nebo, TV, Tintic Valley. Other features mentioned in the text: Wa, Wasatch fault; PC, Parleys Canyon syncline; U, Uinta arch; GSL, Great Salt Lake; SLC, Salt Lake City; T, Tooele.

Christie-Blick and Link, 1988; Christie-Blick et al., 1989), and they overlie approximately 5 km of shallow marine quartzite and siltstone of late Mesoproterozoic to early Neoproterozoic age (Big Cottonwood Formation) not exposed in the Sheeprock Mountains (Fig. 2; Link et al., 1993; Ehlers et al., 1997). The Sheeprock Mountains locality is amongst the first at which sequence stratigraphic principles were applied to Proterozoic rocks (Christie-Blick et al., 1988, 1995; Christie-Blick and Levy, 1989b; Levy and Christie-Blick, 1991a; Levy et al., 1994). Unconformity-bounded sequences recognized there can be traced into the thinner Wasatch Range succession, providing insights about stratigraphic relations not resolved in earlier geological mapping (Christie-Blick, 1982; Levy and Christie-Blick, 1991a; Levy et al., 1994).

Day 1 (Friday, 17 October, 1997). Day 1 is devoted to a traverse at the northern flank of the Sheeprock Mountains (Harker Canyon; HC in Fig. 3) from the base of the

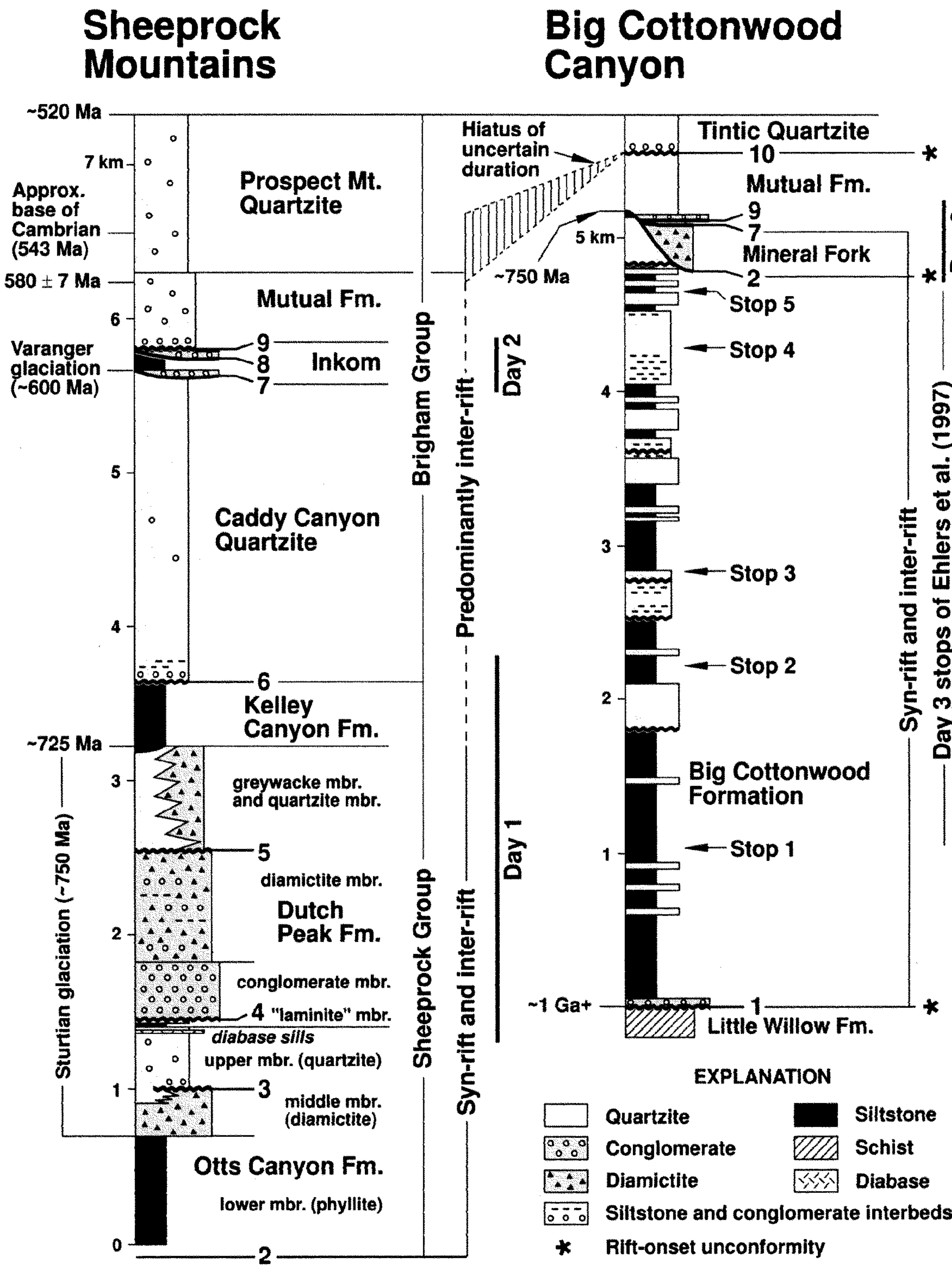

Figure 2. Proterozoic and Lower Cambrian stratigraphy of the Sheeprock Mountains and Big Cottonwood Canyon, and stratigraphic focus of field excursion traverses and stops, with an interpretation of age, relation to Sturtian and Varanger glaciation, location of sequence boundaries (informally numbered) and tectonic setting. Rift-onset unconformities are interpreted at three levels. Stratigraphic data for the Big Cottonwood Formation are mostly from Ehlers et al. (1997).

glacial-marine Dutch Peak Formation, through the Kelley Canyon Formation (basinal siltstone), to the basal part of the Caddy Canyon Quartzite (braided-fluvial pebbly quartzite and siltstone). The rocks dip steeply towards the north, so that the oldest strata are exposed at the highest elevations (in the vicinity of Dutch Peak; 2732 m or 8964 ft). Also preserved near the top of Dutch Peak is some of the best evidence for glaciation in the Sheeprock Mountains (dropstones and a striated clast). Participants should be prepared for a hike up a steep ridge with an elevation change of about 650 m (2150 ft).

Day 2 (Saturday, 18 October, 1997). The focus of Day 2 is the sedimentology and sequence stratigraphy of the interval from the upper part of the Caddy Canyon Quartzite, through the Inkom Formation (braided-fluvial conglomerate and marine siltstone), to the basal part of the Mutual Formation (also braided-fluvial conglomerate). These rocks are well exposed in an overturned structural panel approximately 3 km (2 miles) southwest of Horse Valley on the southern side of the Sheeprock Mountains (HV in Fig. 3). Sequence boundaries with incised valleys 30 to 160 m deep are present at three levels in that area: at the top of the Caddy Canyon Quartzite (the deepest valley), within the Inkom Formation and at the base of the Mutual Formation. Valley fills of Proterozoic age tend to be rather coarser-grained than their Phanerozoic counterparts, and these examples are especially impressive, with outsize intraformational argillite clasts up to 3 m across. The surface at the top of the Caddy Canyon Quartzite is thought to correspond with glacial-eustatic drawdown at the onset of Varanger glaciation (Levy and Christie-Blick, 1991a; Link et al., 1993; Levy et al., 1994), although direct evidence for a second glaciation has been recognized in the Cordillera only in western Canada (Aitken, 1991; Hein and McMechan, 1994; Ross et al., 1995). The traverse is over rolling topography at an elevation of about 2,125 to 2,250 m (7,000 to 7,400 ft), and approximately 5 to 7 km (3 to 4 miles) for the round trip, depending on how closely we can approach the outcrops with available vehicles.

Day 3 (Sunday, 19 October, 1997). The final day of the excursion is divided into two parts. The first part is coordinated with a one-day excursion to the Big Cottonwood Formation of the central Wasatch Range (Fig. 1; and Fig. 2 of Ehlers et al., 1997), an opportunity to examine some remarkable quartzite-siltstone tidal rhythmites. All stops are close to the road in Big Cottonwood Canyon. The second part involves a traverse at Mill B North Fork of Big Cottonwood Canyon from the upper part of the Big Cottonwood Formation through the glacial-marine Mineral Fork Formation to the lower part of the Mutual Formation (mostly braided-fluvial conglomerate and quartzite). Sequence stratigraphic studies indicate that the sequence boundary at the base of the Mutual corresponds in the southern Sheeprock Mountains to the top of the Caddy Canyon Quartzite. Strata equivalent to much of the Kelley Canyon Formation and Caddy Canyon Quartzite (about 2,500 m thick in the Sheeprock Mountains) are not represented in Big Cottonwood Canyon. The main outcrops in Mill B North Fork, located at an elevation of 2,250 to 2,375 m (7,400 to 7,800 ft), are reached up a steep switchback trail, with an elevation change of about 425 m (1,400 ft). Agile participants can reach the basal contact of the Mineral Fork Formation approximately 500 m (1,650 ft) west-northwest of the trail, where glacial grooves of Proterozoic age are preserved on the exhumed unconformity surface.

STRATIGRAPHIC AND TECTONIC FRAMEWORK

SHEEPROCK MOUNTAINS

The Sheeprock Mountains (Fig. 1) are underlain by an allochthonous succession of little-metamorphosed Neoproterozoic to Lower Cambrian clastic sedimentary rocks and Paleozoic carbonate rocks with an aggregrate thickness of more than 12,500 m (Cohenour, 1959; Groff, 1959; Christie-Blick, 1982, 1983a; Hintze, 1988). These were deformed, presumably in late Cretaceous time, and transported tectonically eastwards from a palinspastic position close to the Utah-Nevada state line (Levy and Christie-Blick, 1989). The southern flank of the range and much of the adjacent West Tintic Mountains are overlain and intruded by post-orogenic volcanic and plutonic rocks of Oligocene to Miocene age (Morris and Kopf, 1967, 1970a, 1970b).

The Neoproterozoic and Lower Cambrian part of the succession begins at the base with 2,700 to 4,300 m of phyllite, quartzite, glacial-marine diamictite and siltstone assigned to the Otts Canyon, Dutch Peak and Kelley Canyon formations of the Sheeprock Group (Fig. 2). These units are overlain by 1,950 to 4,000 m of quartzite and minor siltstone assigned to the Caddy Canyon Quartzite, Inkom and Mutual formations and Prospect Mountain Quartzite, which together constitute the Brigham Group. Descriptions of each of these units and information about regional and local lateral variations in thickness and facies may be found in Blick (1979), Christie-Blick (1982), Crittenden et al. (1983), Christie-Blick and Levy (1989a), Levy and Christie-Blick (1991a), and Link et al. (1993, 1994).

The structure of the Sheeprock Mountains is dominated by the Sheeprock and Pole Canyon thrusts of presumed late Cretaceous age, and along the northern flank of the range by the Harker, Lion Hill and related low-angle faults (Fig. 3). These latter faults place younger on

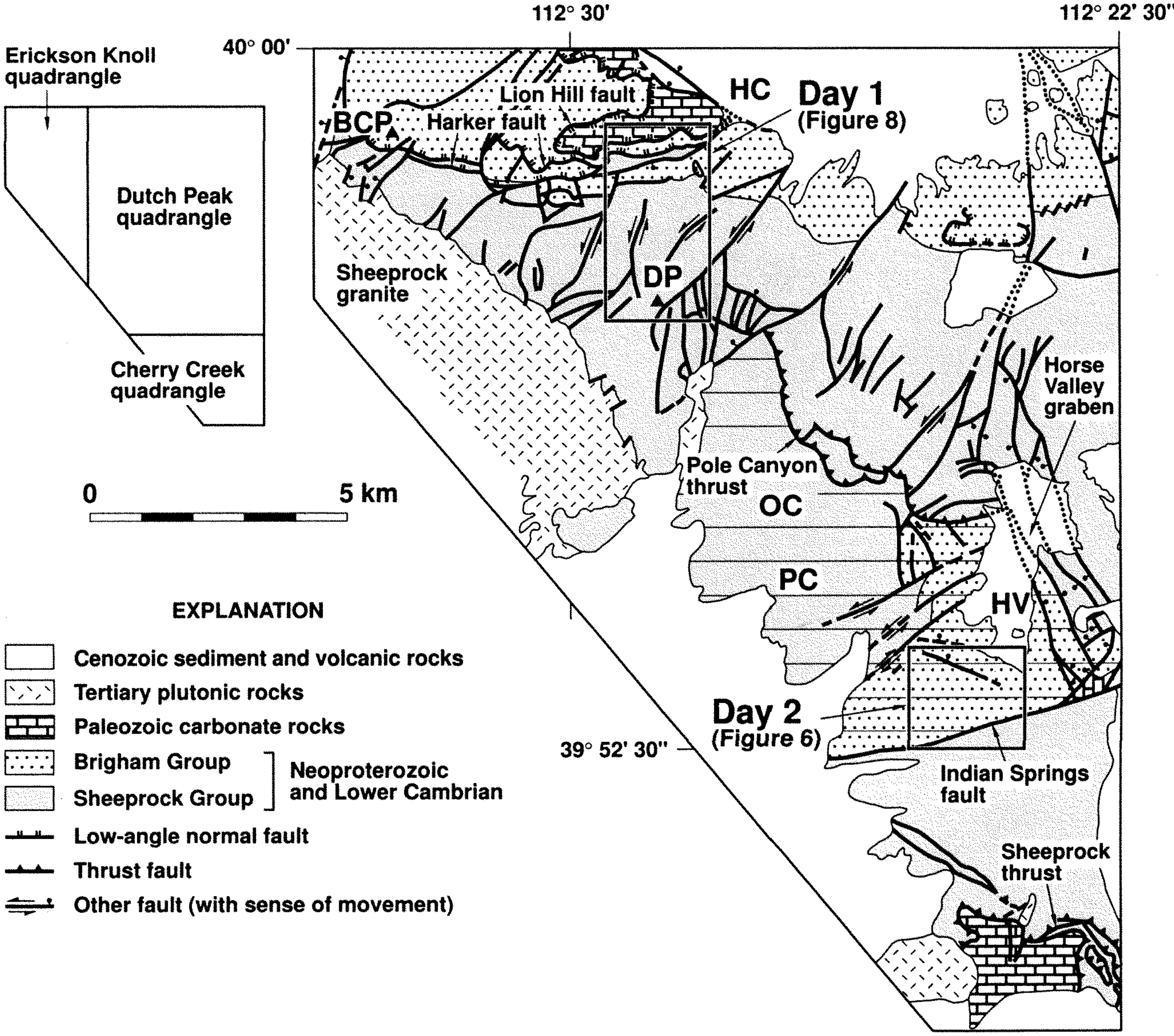

Figure 3. Simplified geological map of the southern part of the Sheeprock Mountains (from Morris and Kopf, 1970a; Blick, 1979; Christie-Blick, 1983a; and unpublished mapping), and location of traverses for Days 1 and 2 of the excursion (see Figs. 6 and 8 below). Horizontal ruling indicates overturned rocks between the Sheeprock and Pole Canyon thrusts. Geographic features: BCP, Black Crook Peak; DP, Dutch Peak; HC, Harker Canyon; HV, Horse Valley; OC, Otts Canyon; PC, Pole Canyon.

older rocks with as much as 3,500 m of stratigraphic omission, and are inferred to be of late Cenozoic age (Christie-Blick, 1983a). Several steeply dipping faults are interpreted as tear faults. Of these, the Indian Springs fault is most conspicuous (Groff, 1959; Morris and Kopf, 1967, 1970a, 1970b). Other high-angle faults (e.g., Horse Valley graben) are probably related to late Cenozoic extension.

The Sheeprock and Pole Canyon thrusts were regarded by Morris (1977, 1983) and Christie-Blick (1983a) as discrete although genetically related structures (Fig. 4). An earlier idea, that they are different segments of the same fault (Armstrong, 1968), has been revived recently by Mukul and Mitra (1994). I continue to prefer the two-fault interpretation because the two structures are characterized by markedly different stratigraphic separation (no more than 6 km for the Pole Canyon thrust, and in excess of 10 km for the Sheeprock thrust). Outcrops for Day 1 are in the upper plate of the Pole Canyon thrust, and those for Day 2 are in overturned strata of the lower plate (Figs. 3 and 4).

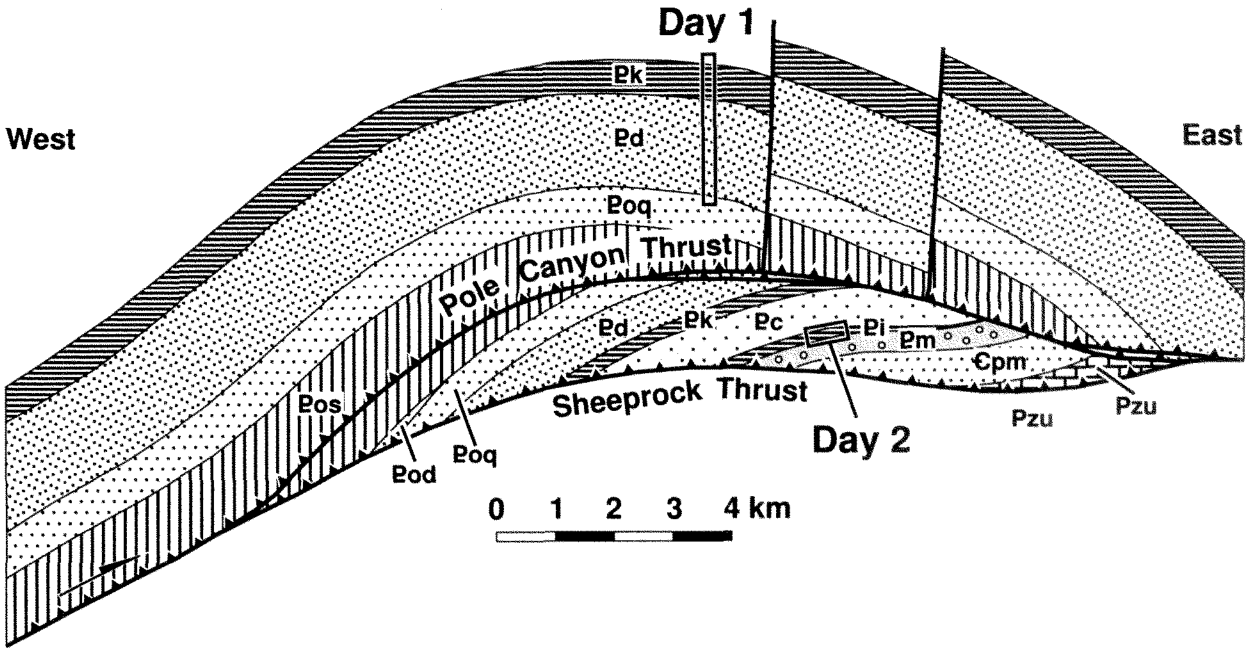

Figure 4. Conceptual cross section showing the interpreted relation between the Sheeprock and Pole Canyon thrusts and tear faults in the upper plate of the Pole Canyon thrust (from Christie-Blick, 1983a), and location of traverses for Days 1 and 2 of the excursion. Cenozoic structure has been omitted for simplicity. Stratigraphic units: Ꝑos, Ꝑod and Ꝑoq, lower (phyllite), middle (diamictite) and upper (quartzite) members of the Otts Canyon Formation; Ꝑd, Dutch Peak Formation; Ꝑk, Kelley Canyon Formation; Ꝑc, Caddy Canyon Quartzite; Ꝑi, Inkom Formation; Ꝑm, Mutual Formation; Ꞓpm, Prospect Mountain Quartzite; Pzu, undifferentiated Paleozoic rocks.

CENTRAL WASATCH RANGE

Proterozoic, Paleozoic and Mesozoic rocks in the vicinity of Big Cottonwood Canyon of the central Wasatch Range are parautochthonous with respect to the continental interior, and they occupy a palinspastic position considerably east of the Sheeprock Mountains (Fig. 1; Crittenden, 1976; Bruhn et al., 1986; Hintze, 1988; Christie-Blick et al., 1989; Levy and Christie-Blick, 1989; Ehlers et al., 1997). The Big Cottonwood Formation, a succession approximately 5 km thick of quartzite and siltstone of late Mesoproterozoic to early Neoproterozoic age, is unconformably overlain in the vicinity of Big Cottonwood Canyon by up to 800 m of glaciogenic diamictite, siltstone, sandstone and conglomerate (Mineral Fork Formation), and in turn by about 600 m of quartzite and conglomerate of the Mutual Formation and Tintic Quartzite (latest Proterozoic to Early Cambrian; Fig. 2; Crittenden et al., 1952; Crittenden, 1976; Lochman-Balk, 1976; Ehlers et al., 1997). The Mineral Fork and overlying units are grossly equivalent to the considerably thicker Sheeprock and Brigham groups respectively of the Sheeprock Mountains.

The rocks are located in the lower part of an impressive homocline of Proterozoic to Mesozoic strata that make up the southern limb of the northeast-trending Parleys Canyon syncline (PC in Fig. 1), and are broken only on the north side of Big Cottonwood Canyon by the Mount Raymond thrust (MR in Fig. 1; Granger and Sharp, 1952; Crittenden, 1976). This fault was interpreted by Bruhn et al. (1986) as a lateral ramp linking the Charleston thrust in the south with the Absaroka thrust in the north (C-N and A in Fig. 1), faults that are located structurally below all of the thick successions of Neoproterozoic strata in western Utah, including the succession in the Sheeprock Mountains. On the northern limb of Parleys Canyon syncline, the Archean and Paleoproterozoic Farmington Canyon Complex is overlain directly by the Cambrian Tintic Quartzite, and the intervening Big Cottonwood, Mineral Fork and Mutual formations are absent (see Fig. 7 below; Crittenden and Wallace, 1973; Lochman-Balk, 1976; Bruhn et al., 1986; Bryant, 1990). Even when allowance is made for eastward displacement of up to 35 km on the Mount Raymond thrust (Levy and Christie-Blick, 1989), the abrupt pinch-out of these units is remarkable. Structurally below the Mount Raymond thrust and disrupted by later faulting, segments of the Alta thrust are present on both sides of Big Cottonwood Canyon at the level of the uppermost Big Cottonwood, Mineral Fork and Mutual formations. This minor splay of the Absaroka-Charleston system is of interest because it repeats the unconformity-bounded panel of Mineral Fork Formation that we shall examine on the last afternoon of the field trip.

SEQUENCE STRATIGRAPHY

Many of the participants in the excursion will be familiar with the fundamentals of sequence stratigraphy, but because this perspective of sedimentary geology underpins much of the interpretation presented, I include here a few summary comments, with connections to the examples to be examined. Sequence stratigraphy makes use of the fact that sedimentary successions are pervaded by physical discontinuities (Vail, 1987; Van Wagoner et al., 1988, 1990; Loucks and Sarg, 1993; Posamentier and James, 1993; Weimer and Posamentier, 1993; Christie-Blick and Driscoll, 1995). These are present at a great range of scales, and they arise in a number of different ways, but they have one important attribute in common. In each case, at least as a first approximation, sediments accumulating above a particular surface are younger than those below the surface. The essence of sequence stratigraphy therefore is the tracing of surfaces, which in many cases can be shown to pass through laterally changing facies and, on a regional scale, from one lithostratigraphic unit to another.

The principal discontinuities in a succession, termed sequence boundaries, are unconformities that are related (or inferred to be related) at least locally to the lowering of depositional base level, and hence to subaerial erosion or sediment bypassing (Christie-Blick and Driscoll, 1995). In this context, evidence for exposure (e.g., a paleosol or vadose diagenesis) is not sufficient for the interpretation of a sequence boundary if it cannot be demonstrated that base level was lowered, and evidence for erosion does not by itself require subaerial exposure. Channels are present locally in a wide range of depositional settings. Sedimentary successions also contain numerous other sharp contacts, many of which are unrelated to sequence boundaries (e.g., marine flooding surfaces, associated with abrupt upward deepening).

Most preserved Proterozoic deposits accumulated in intracratonic and passive-margin settings characterized by shallow depositional ramps lacking pronounced breaks in slope. In the absence of syndepositional faulting, Proterozoic stratigraphy tends to be relatively persistent and apparently conformable over large areas. The key to sequence stratigraphy in these rocks is the identification of incised valleys (Christie-Blick et al., 1995). These are associated with mappable surfaces (sequence boundaries), and the erosional relief of valleys (tens of metres) is commonly greater than that of channels within the same succession. Valleys are typically filled by a combination of fluvial and braid-deltaic to shallow marine deposits, which in many Proterozoic examples are unusually coarse-grained, consistent with an abrupt upward shoaling and/or basinward shift of sedimentary facies across the boundary. Valley walls in some cases exhibit evidence for a hiatus in sedimentation between erosion and subsequent deposition (e.g., evidence that the wall rocks were already consolidated or lithified, or the presence of paleosols). Unlike most channels, sequence boundaries are also located systematically where there is a change in the stacking pattern of small-scale shoaling-upward successions (parasequences) from a forestepping to a backstepping motif. Examples of valleys will be demonstrated during the excursion at three levels within the Brigham Group, and at the base of the glaciogenic Mineral Fork Formation where they are as much as 900 m deep (Days 2 and 3). Incised valleys may also be present in the Big Cottonwood Formation (Ehlers et al., 1997), although that interpretation has not yet been tested by the lateral tracing of surfaces.

Sequence boundaries may be identifiable also in strata that accumulated in somewhat deeper water, especially where the existence of paleoslopes and/or syndepositional tilting resulted in the development of sedimentary wedges and accentuated offlap and onlap geometry. [The terms offlap and onlap refer to the progressive up-dip or lateral termination of strata against an overlying and underlying surface, respectively (Mitchum, 1977; Christie-Blick, 1991).] Evidence for erosion is present locally, both up dip as a result of wave or current activity on the shallow shelf, and down dip as a result of mass wasting; and a discontinuity may be recognizable also by the superposition of contrasting facies. Possible examples in a glacial and glacial-marine setting will be pointed out during the excursion in the Sheeprock Group of the Sheeprock Mountains (Day 1), although definitive evidence is difficult to establish in available outcrop.

A general issue that will probably arise during the excursion concerns the degree of conformability in successions containing sequence boundaries. Sequence boundaries are obviously discontinuities, but their presence does not necessarily imply the existence of pronounced breaks in sedimentation. For example, seismic stratigraphic studies across the New Jersey continental margin demonstrate the existence of at least 12 sequence boundaries in the lower and middle Miocene, in a span of only 13 m.y. (Miller et al., 1996; N. Christie-Blick et al., unpublished data). On the other hand, stratal concordance is no guarantee of conformability. North of Big Cottonwood Canyon, the contact between the Big Cottonwood Formation and Mutual Formation is remarkably concordant, and were it not for a marked color difference in otherwise similar facies and the erosional relief locally developed at the base of the intervening Mineral Fork Formation, the existence of a hiatus of up to several hundred million years might be missed entirely (Crittenden et al., 1952).

Most thick Proterozoic successions are deformed to some extent, and there are commonly limitations to the distance over which a given sequence boundary can be traced continuously and to the confidence with which a boundary can be correlated between isolated outcrops in the absence of precise biostratigraphy. This means that regional interpretations of Proterozoic stratigraphy inevitably focus on the most prominent surfaces, and that some sequence boundaries may be recognizable only locally. The tracing of physical surfaces nevertheless provides a useful stratigraphic framework in many Proterozoic successions, and at a range of scales, not in spite of limitations in biostratigraphic and other stratigraphic techniques, but perhaps because of them!

The sequence stratigraphy and sedimentology of the Brigham Group are discussed in some detail in Levy (1991), Levy and Christie-Blick (1991a) and Levy et al. (1994). Here I draw attention to previously unrecognized details from the Brigham Group of the southern Sheeprock Mountains (Day 2), and outline some tentative new ideas about the sequence stratigraphy of the Sheeprock Group (Day 1), in the context of earlier descriptions and interpretations (Blick, 1979; Christie-Blick, 1982; Crittenden et al., 1983; Link et al., 1994). More complete development of the interpretation for the Sheeprock Group will be published elsewhere. Sequence boundaries are numbered informally for the sake of clarity (Fig. 2).

SHEEPROCK GROUP

Geological mapping in the late 1970s (Blick, 1979; Christie-Blick, 1982) revealed some puzzling lateral variations in thickness and facies in both the Otts Canyon and Dutch Peak formations of the Sheeprock Group. The most pronounced variations, along the northern flank of the Sheeprock Mountains (Fig. 5), are in an area of considerable structural complexity as well as incomplete outcrop. The most reasonable interpretation at the time was that the stratigraphic variations are due to lateral intertonguing of sedimentary facies at the margin of a differentially subsiding basin (Fig. 5A; Christie-Blick, 1982; Crittenden et al., 1983). Such intertonguing is demonstrable at the top of the Dutch Peak Formation, between the quartzite and greywacke members and, in the lower plate of the Pole Canyon thrust on the southern flank of the range, between the upper (quartzite) member of the Otts Canyon Formation and the underlying middle (diamictite) member (in that area up to 500 m thick).

The Sheeprock Group is composed primarily of compositionally and texturally heterogeneous stratified diamictite, with several hundred metres of phyllite at the base and siltstone at the top. In the absence of shallow-water indicators, or of firm facies evidence that the ice sheet was ever grounded in this area, the rocks were interpreted to have accumulated in relatively deep water. The intervals of quartzite, as much as several hundred metres thick in both the Otts Canyon and Dutch Peak formations, were viewed as representing somewhat shallower marine conditions, or at least derivation from a shallower water setting, but no good explanation was found for either their compositional and textural maturity or for the process by which they might have accumulated. The quartzites are commonly laminated, rarely cross-stratified, and have none of the obvious features of mass-flow deposits (e.g., sole marks, grading, and dish-and-pillar and other dewatering structures).

Subsequently (in Link et al., 1994), I proposed that the quartzites might have accumulated in a glacial-fluvial to braid-deltaic setting, and that the overall stratigraphic relations are consistent with two large-scale shoaling-upward successions, one corresponding approximately with the Otts Canyon Formation, and the other with the Dutch Peak Formation. Modification to that interpretation is now suggested by the recognition of three stratigraphic discontinuities, here interpreted as sequence boundaries (Fig. 5B). The discontinuities are present at or near the base of the upper (quartzite) member of the Otts Canyon Formation, at or near the base of the conglomerate member of the Dutch Peak Formation, and at the base of the quartzite member of the Dutch Peak Formation (sequence boundaries 3 to 5 in Figs. 2 and 5B).

Upper Otts Canyon sequence boundary. On the northern flank of the Sheeprock Mountains, the middle (diamictite) member of the Otts Canyon Formation is less than 70 m thick, and pinches out towards the northwest. A thin but mappable unit of pebble and granule conglomerate at or near the base of the upper (quartzite) member is composed mainly of extrabasinal clasts, but it contains accessory fragments as large as 50 cm of diamictite and argillite derived from underlying beds. The conglomerate is typically 2–3 m thick; locally it fills channels as deep as 20 m. Where diamictite is absent, quartzite or conglomerate rests with sharp contact on phyllite of the lower member. These observations are consistent with erosional truncation towards the basin margin. The interpreted sequence boundary (surface 3 in Fig. 5B) has not yet been located in the lower plate of the Pole Canyon thrust.

Dutch Peak conglomerate sequence boundary. The second contact, in the lower part of the Dutch Peak Formation

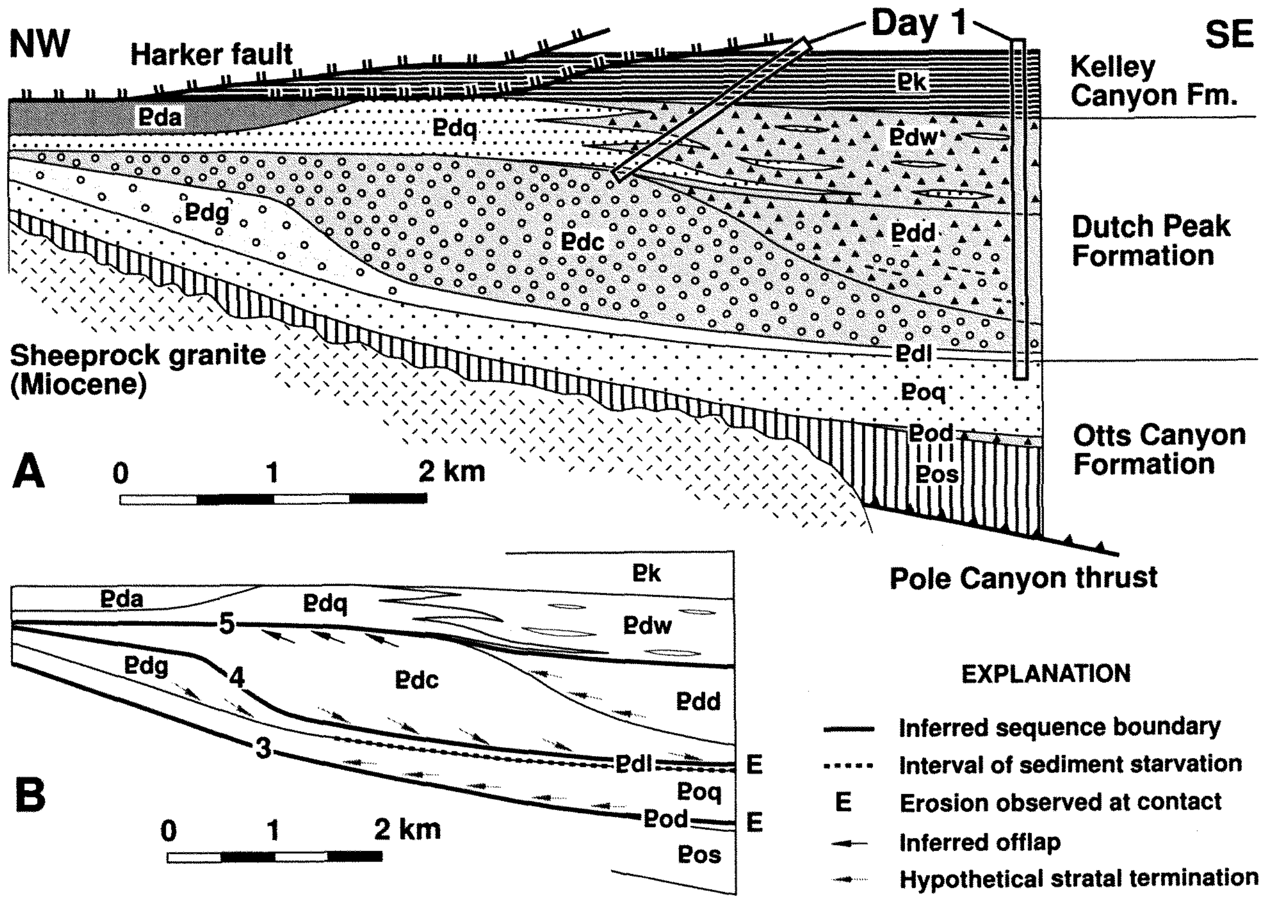

Figure 5. Lateral variations in thickness and facies in the Otts Canyon and Dutch Peak formations on the northern flank of the Sheeprock Mountains, between Black Crook Peak (NW) and Dutch Peak (SE), in the upper plate of the Pole Canyon thrust (A, modified from Christie-Blick, 1982); and a revised interpretation (B) that recognizes the existence of stratigraphic discontinuities (sequence boundaries) within the succession. Stratigraphic units: Ᵽos, Ᵽod and Ᵽoq, lower (phyllite), middle (diamictite) and upper (quartzite) members of the Otts Canyon Formation; Ᵽdl, Ᵽdc, Ᵽdd, Ᵽdw, Ᵽdg, Ᵽdq and Ᵽda, "laminite", conglomerate, diamictite, greywacke, "grit", quartzite and argillite members of the Dutch Peak Formation; Ᵽk, Kelley Canyon Formation. Surfaces cannot be traced continuously along this profile owing to structural complexity and incomplete exposure. Vertical and horizontal scales are equal.

(surface 4 in Fig. 5B), is marked at Dutch Peak by an abrupt upward change from medium- to fine-grained quartzite to disorganized boulder conglomerate containing an anomalous 17% of clasts of diabase similar to that intruding the uppermost part of the Otts Canyon Formation. This contact may trace northwestward, towards the basin margin, to a position at or near the top of the locally developed "grit" member of the Dutch Peak Formation (granule-bearing quartzite, with minor conglomerate, sandy diamictite and sandstone). Through much of the Sheeprock Mountains, the base of the Dutch Peak Formation is marked by a distinctive unit, a few metres to 50 m thick, of finely laminated siltstone and fine- to very fine-grained metallic bluish-grey sandstone, in places containing good examples of dropstones (ice-rafted debris). This unit (informally, the "laminite" member) pinches out towards the basin margin, and is now best interpreted as an interval of sediment starvation (condensed section) within the sequence that corresponds approximately with the upper member of the Otts Canyon Formation and the "grit" member of the Dutch Peak Formation.

Dutch Peak quartzite sequence boundary. The third contact (surface 5 in Fig. 5B), at the base of the quartzite member of the Dutch Peak Formation, appears to truncate bedding in the underlying conglomerate member towards the northwest. The contact traces basinward (towards the southeast) below the greywacke member. That unit contains abundant quartzite lenses, and these are thickest and most continuous in the lower part of the member, consistent with a backstepping stratigraphic motif and overall glacial retreat.

Abrupt thinning of the diamictite member of the Dutch Peak Formation towards the northwest may be interpreted in two ways. One is that it intertongues with the conglomerate member, as originally thought (Christie-Blick, 1982). An alternative, consistent with the lack of evidence for appreciable intertonguing, is that the contact is at least in part a surface of marine onlap (Fig. 5B). In that case, the overall wedge-shaped geometry of the conglomerate member, along with possible offlap at its upper contact (beneath the quartzite member), suggests deposition as an ice-proximal subaqueous fan during glacial advance. The diamictite member is interpreted to have accumulated in a glacial-marine environment, during glacial retreat, from a combination of ice-rafting, sediment gravity flow and winnowing by bottom currents (Blick, 1979; Crittenden et al., 1983; Link et al., 1994). The sharp contact at the base of the quartzite member of the Dutch Peak Formation is tentatively attributed to a combination of grounding of the ice sheet below sea level and isostatic rebound during glacial retreat. Upward fining at the top of the Dutch Peak Formation, from sandy diamictite into siltstone of the Kelley Canyon Formation, is thought to be due to a reduction in the supply of terrigenous sediment, perhaps combined with a glacial-eustatic rise.

If these tentative new ideas about the sequence stratigraphy of the Sheeprock Group are correct, they are significant in several respects. It is implied that the ice sheet was grounded farther west than previously supposed (Blick, 1979; Crittenden et al., 1983). If the contact between the conglomerate and diamictite members is an onlap surface, rather than due to an abrupt lateral change in facies, complementary variations in the thicknesses of these units are a measure of paleobathymetric relief. Paleowater depths may have been as great as several hundred metres. This is large, but not unreasonable in a tectonically active basin and in the vicinity of a grounded ice sheet. The Antarctic ice sheet is today grounded at depths as great as 1,300 m below sea level (Denton et al., 1971; Crabtree, 1981). Water depths of hundreds of metres are also inferred for the correlative Mineral Fork Formation of the central Wasatch Range (Christie-Blick, 1983b). Finally, the sequence stratigraphic context of the quartzite unit in the Dutch Peak Formation suggests an origin as ice-proximal glacial-marine outwash. Its compositional and textural maturity, so markedly in contrast to that of associated facies, indicates a local provenance perhaps related to glacial erosion of older Proterozoic quartzites such as the Big Cottonwood Formation, as is demonstrably the case for the Mineral Fork Formation.

BRIGHAM GROUP

Regional sequence boundaries have been identified at three stratigraphic levels in the Brigham Group of northern and western Utah and adjacent Idaho: 1) at or near the top of the Caddy Canyon Quartzite, 2) at or near the base of the Mutual Formation, and 3) at the base of the Tintic, Geertsen Canyon and Camelback Mountain quartzites (sequence boundaries 7, 9 and 10 in Fig. 2; Christie-Blick and Levy, 1985, 1989a, 1989b; Link et al., 1987; Christie-Blick et al., 1988; Levy, 1991; Levy and Christie-Blick, 1991a; Link et al., 1993; Levy et al., 1994). Of these, the upper Caddy Canyon and base-Mutual sequence boundaries can be identified with the greatest confidence and traced over a distance of several hundred kilometres. The base-Tintic sequence boundary is confidently identified only in northern Utah and southeastern Idaho; elsewhere the correlative surface is cryptic. One to several higher-order sequence boundaries are present locally within the Inkom Formation (e.g., surface 8 in Fig. 2), but none of these can be correlated with confidence between mountain ranges. Examples of all of these surfaces will be examined during the excursion.

The principal evidence for the existence of sequence boundaries is the presence at several localities of incised valleys, up to several tens of metres deep, and associated with marked upward coarsening and/or shoaling of sedimentary facies (Dugway Range, Sheeprock Mountains, Canyon Range, Huntsville, Portneuf Range and Big Cottonwood Canyon of the central Wasatch Range; Fig. 1; Levy et al., 1994). At other localities (e.g., Drum Mountains and Bannock Range; Fig. 1), limited outcrop continuity makes it difficult to recognize erosional relief, but the presence of facies discontinuities nevertheless permits sequence boundaries to be interpreted within the succession.

Recent work near Horse Valley in the southern Sheeprock Mountains (HV in Fig. 3) has shown that erosional relief at the upper Caddy Canyon sequence boundary may be as large as 160 m, considerably greater than the 45 m estimated by Levy et al. (1994) at this locality. It has also been possible to demonstrate, for the first time in the miogeoclinal succession, more than 30 m of erosional relief at the base-Mutual sequence boundary. These conclusions are based on revision of existing geological mapping (Morris and Kopf, 1970a; Blick, 1979; Christie-Blick. 1983a), and on the measurement of a series of new sections through the Inkom Formation and basal part of the Mutual Formation (Fig. 6).

Marjorie Levy and I recognized more than 10 years ago that, in the vicinity of sections c to g (Fig. 6A), the contact mapped by Morris and Kopf (1970a) as the base of the Mutual Formation is actually a sequence boundary

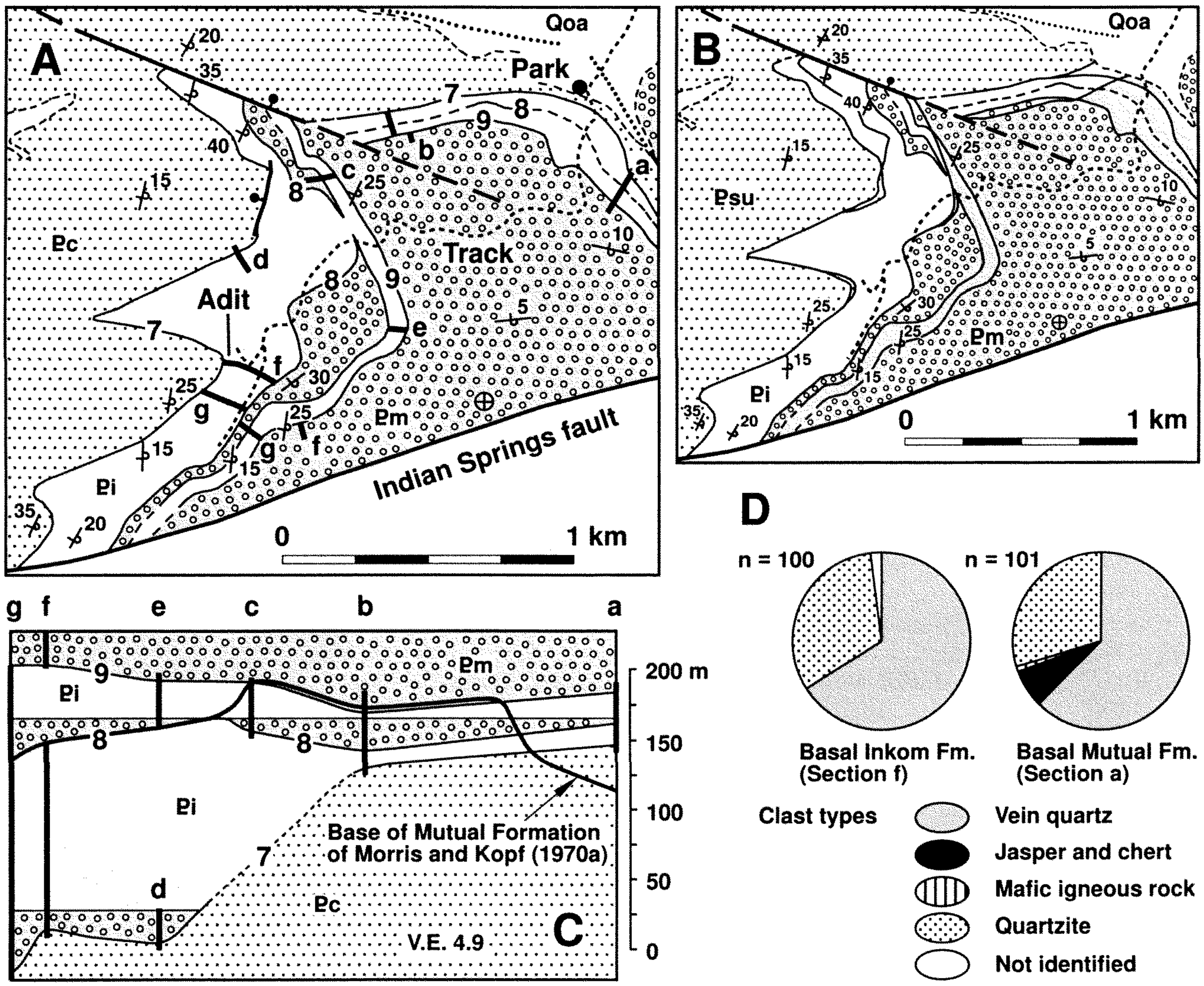

Figure 6. A. Geological map of the Horse Valley locality, Sheeprock Mountains (modified from Morris and Kopf, 1970a; Blick, 1979; Christie-Blick, 1982), showing location of sequence boundaries 7 to 9 (see Fig. 2) and stratigraphic sections a to g (see Fig. 12 below). See Figure 3 for location of the map area (Day 2) and standard geological symbols. B. Geological map of the same area by Morris and Kopf (1970a), with contacts from panel A shown for reference. Note different interpretations in the northern part of the area at the contacts between the Caddy Canyon Quartzite (Ꝑc), Inkom Formation (Ꝑi) and Mutual Formation (Ꝑm). Rocks now assigned to the Caddy Canyon Quartzite were included by Morris and Kopf (1970a) in the upper member of the Sheeprock Formation (Psu). Qoa is older Quaternary alluvium. C. Simplified stratigraphic cross section showing incised valleys at sequence boundaries 7 to 9, and the mapped location of the base of the Mutual Formation from Morris and Kopf (1970a). Vertical exaggeration is 4.9. D. Clast counts (pie diagrams) for conglomerate units in the Inkom and Mutual formations. Minimum clast size counted, 1 cm.

within the Inkom Formation (Fig. 6B and C). However, we paid less attention to the panel of less well exposed Inkom Formation to the north (at section b in Fig. 6A), which we regarded as anomalously thin, possibly as a result of tectonic flattening beneath the Pole Canyon thrust. New mapping shows that Inkom stratigraphy in that area is consistent with that to the south, and that it is the lower part of the Inkom that is thin (Fig. 6C). This is important because it suggests that the change in thickness is not due primarily to deformation. Moreover, details of the stratigraphy initially appeared anomalous in limited outcrop at least in part as a result of mismapping of the Inkom Formation east of section b. Near section b, the contact mapped by Morris and Kopf as the base of the Mutual Formation is close to the base-Mutual sequence boundary of Levy et al. (1994). However, Morris and Kopf traced this contact eastward and connected it with an outcrop of conglomerate that they interpreted as Mutual, but which is actually basal Inkom (Fig. 6B). The effect of this interpretation was to make the "Inkom Formation" appear unusually coarse-grained because, to the east of section b, the rocks mapped as Inkom by Morris and Kopf are in part Caddy Canyon Quartzite. In fact, the Inkom outcrop turns abruptly southward at section a, where the stratigraphy is similar to that at section c (Fig. 6C). Abrupt southward thickening of the lower Inkom between sections b and g is attributed to erosional relief on the underlying sequence boundary, relief that is considerably greater than the thickness of conglomerate at the base of the Inkom Formation in this area (and the basis for the estimate of Levy et al., 1994). The possibility that the observed thickening is due to the presence of a growth fault is regarded as unlikely because such a structure would need to have been active only during deposition of the Inkom Formation, and such faults are preserved nowhere else at this stratigraphic level.

Erosional relief at the sequence boundaries within the Inkom Formation and at the base of the Mutual Formation can similarly be monitored with reference to the flooding surface at the top of the upper Inkom conglomerate (Fig. 6C). This conglomerate ranges in thickness from 0 to about 30 m, which can be taken as a minimum estimate for depth of valleys at that level. The upper siltstone unit of the Inkom is more than 37 m thick at section g, and 18 m thick at section a, but only 4 m thick at section b, consistent with erosional relief at the base of the Mutual Formation in excess of 30 m. Erosional relief is in most places difficult to document at this contact owing to the absence of suitable markers in either the Inkom Formation or Mutual Formation.

RELATION TO VARANGER GLACIATION

Levy and Christie-Blick (1991a) and Levy et al. (1994) offered three arguments for relating the upper Caddy Canyon sequence boundary to the onset of Varanger glaciation: 1) The stratigraphic level of this surface, with respect to Sturtian-age glacial rocks, appears to be the same as the glaciogenic Ice Brook Formation in the Mackenzie Mountains of northwestern Canada (Aitken, 1991; Ross et al., 1995). 2) Incised valleys are best developed at this horizon, consistent with the relatively large-amplitude sea-level change expected for glacial-eustatic fluctuation. 3) High-order valleys described at several levels in the uppermost Caddy Canyon Quartzite and Inkom Formation are consistent with high-frequency changes in sea level expected during deglaciation. The considerable depth of valleys now documented at the upper Caddy Canyon sequence boundary in the Sheeprock Mountains reinforces this interpretation.

The origin of the base-Mutual sequence boundary is uncertain. Levy and Christie-Blick (1991a) suggested that it might be due to either a glacial-eustatic fall or regional uplift related to the onset of terminal Proterozoic lithospheric extension. Both hypotheses are permitted by an Ar^{40}/Ar^{39} age on hornblende from an alkali trachyte in the Browns Hole Formation at Huntsville, which dates the surface as older than 580±7 Ma (Crittenden and Wallace, 1973; Christie-Blick and Levy, 1989a), but not necessarily much older. The main difficulty with the tectonic hypothesis is that while regional uplift accounts for the abrupt increase in grain size observed over many hundreds of kilometres, such uplift would also reduce the available sedimentary accommodation. Yet the Mutual is typically several hundred metres thick, even in the vicinity of Big Cottonwood Canyon, palinspastically located more than 150 km east of the Neoproterozoic depocenter (Fig. 2; Christie-Blick, 1982; Christie-Blick and Levy, 1989a; Levy and Christie-Blick, 1989). The main difficulty with the glacial-eustatic hypothesis is that while renewed post-glacial eustatic rise accounts for the relatively uniform thickness of the Mutual over a broad area, it does not explain why fluvial sedimentation persisted. One possible explanation is that glacial-eustatic change was accompanied by a marked increase in sediment flux related to either glacial-isostatic rebound or tectonic uplift far from the depocenter. North America was located at mid- to high paleolatitude at the time (Torsvik et al., 1996), and deglaciation would have been delayed with respect to continents such as Australia that were glaciated at lower latitude (Schmidt and Williams, 1995). The abundance of quartzite in the Brigham Group suggests that the North American continent stood high for a considerable time during the late Neoproterozoic and Early Cambrian.

PROTEROZOIC AND CAMBRIAN TECTONICS

Stratigraphic evidence for rifting will be noted during the excursion at three levels (Fig. 2): in the Big Cottonwood Formation of the central Wasatch Range; in the Sheeprock Group of the Sheeprock Mountains; and at the base of the Tintic Quartzite (also in the Wasatch Range). Since Neoproterozoic and lower Paleozoic strata of the North American Cordillera were first interpreted, 25 years ago, as a remnant of an ancient passive continental margin (Stewart, 1972), there has been considerable debate about the timing of continental separation. Apart from uncertainty in the ages of the rocks themselves, this is largely because continental rifting does not necessarily lead to passive-margin development, and many passive margins are underlain by extensional basins of a range of ages.

One school of thought has been to favor continental separation shortly after deposition of the Sturtian glacial strata (approx. 725 Ma in Fig. 2; Stewart, 1972, 1976; 1982; Stewart and Suczek, 1977; Link, 1984; Link et al., 1987; Ross, 1991). Another has been to favor the terminal Proterozoic to early Cambrian timing suggested by quantitative analysis of tectonic subsidence recorded by lower Paleozoic carbonate rocks (Armin and Mayer, 1983; Bond et al., 1983, 1985; Bond and Kominz, 1984; Levy and Christie-Blick, 1991b; Levy et al., 1994). With a correction for recent work in Cambrian geochronology (Grotzinger et al., 1995), this event now is constrained as younger than 570 Ma, and probably younger than 543 Ma (Fig. 2). In the absence of other data, the second option has been the more parsimonious one, but it has never been possible to exclude an earlier (Sturtian) continental separation on the basis of the subsidence analysis. In contrast with my earlier papers, I currently favor the idea that a passive continental margin first developed at about 725 Ma because this is required by the hypothesis that the continental counterpart to Laurentia was eastern Australia and adjacent Antarctica (Hoffman, 1991; Powell et al., 1994; Christie-Blick et al., 1995; Ross et al., 1995; cf. Veevers et al., 1997). If that interpretation is correct, it is necessary to conclude that in the western United States the continental margin was modified by renewed extension in latest Proterozoic and early Cambrian time, in order to account for the very marked acceleration of subsidence observed in the Early Cambrian.

SHEEPROCK GROUP

Abrupt lateral changes in thickness are evident in both the Otts Canyon and Dutch Peak formations in the Sheeprock Mountains (e.g., Fig. 5). These are due in part to variations in paleobathymetry and to erosional truncation at sequence boundaries, as outlined above, and in places to later deformation. However, certain gross patterns are interpreted to reflect syndepositional tilting towards one or more extensional growth faults (not identified in available outcrop). This interpretation is supported by the presence of syndepositional mafic sills in the upper part of the Otts Canyon Formation (Fig. 2; Christie-Blick, 1982; Crittenden et al., 1983; Christie-Blick and Levy, 1989a). Chemical analyses of volcanic rocks at about the same stratigraphic level elsewhere in northern Utah and southeastern Idaho indicate tholeiitic and alkalic affinities consistent with a within-plate extensional tectonic setting (Harper and Link, 1986).

In the upper plate of the Pole Canyon thrust, the upper (quartzite) member of the Otts Canyon Formation thickens to the southeast from 240 m near Black Crook Peak to about 1,000 m east of Horse Valley (BCP and HV in Fig. 3), a distance of approximately 17 km. In the lower plate, it *thins* southward from more than 1,170 m at the head of Otts Canyon to 180 m where the member crosses Pole Canyon only 2.5 km away (OC and PC in Fig. 3). Part of that thickness change is due to interfingering with the middle (diamictite) member of the Otts Canyon Formation, which thickens substantially (to 500 m) in the same direction, and to possible structural complications on the west side of Pole Canyon. However, a significant portion of the thickness variations in the upper member in both thrust sheets must be due to differential subsidence, even though the apparent direction of thickening differs. The Dutch Peak Formation thickens to the southeast from about 650 m near Black Crook Peak to about 1,750 m at its type section north of Dutch Peak (BCP and DP in Fig. 3), a distance of about 7 km. In the lower plate of the Pole Canyon thrust, it thickens southward from 550 m to 750 m in the vicinity of Pole Canyon, a relatively minor change and a trend that differs from that of the Otts Canyon Formation.

BIG COTTONWOOD FORMATION

The present outcrop pattern of the Big Cottonwood Formation and possibly coeval Uinta Mountain Group of the Uinta Mountains to the east (U in Fig. 1) is related to late Mesozoic and early Cenozoic folding, thrusting and uplift, but the geometry of these younger structures is thought to have been influenced strongly by the geometry of the depositional basin (Bruhn et al., 1986; Christie-Blick and Levy, 1989a). Structural and metamorphic studies in the northeastern part of the Uinta Mountains suggest that Archean and Paleoproterozoic crystalline rocks were subject to block faulting during the Mesoproterozoic, and that the Uinta Mountain Group accumulated in an east-trending graben or half-graben (Sears et al., 1982). The northward pinch-out of the Big Cottonwood Formation across Parleys Canyon syncline is consistent with this interpretation (Fig. 7; Crittenden and Wallace, 1973; Bruhn et al., 1986; Christie-Blick and Levy, 1989a; Ehlers et al., 1997). The possibility that the thickness change is due largely to subsequent erosion is not supported by studies of clast composition in Neoproterozoic glacial deposits (Blick, 1979; Crittenden et al., 1983; Christie-Blick and Levy, 1989a). These studies show that in the region north of the present outcrop of the Big Cottonwood Formation, crystalline basement was already widely exposed at the onset of glaciation.

In their companion excursion guide, Ehlers et al. (1997) raise the possibility that the Big Cottonwood Formation might be sufficiently young to relate directly to the initial separation event (725 Ma in Fig. 2). Available geochronology places only broad limits on the age of the rocks, and

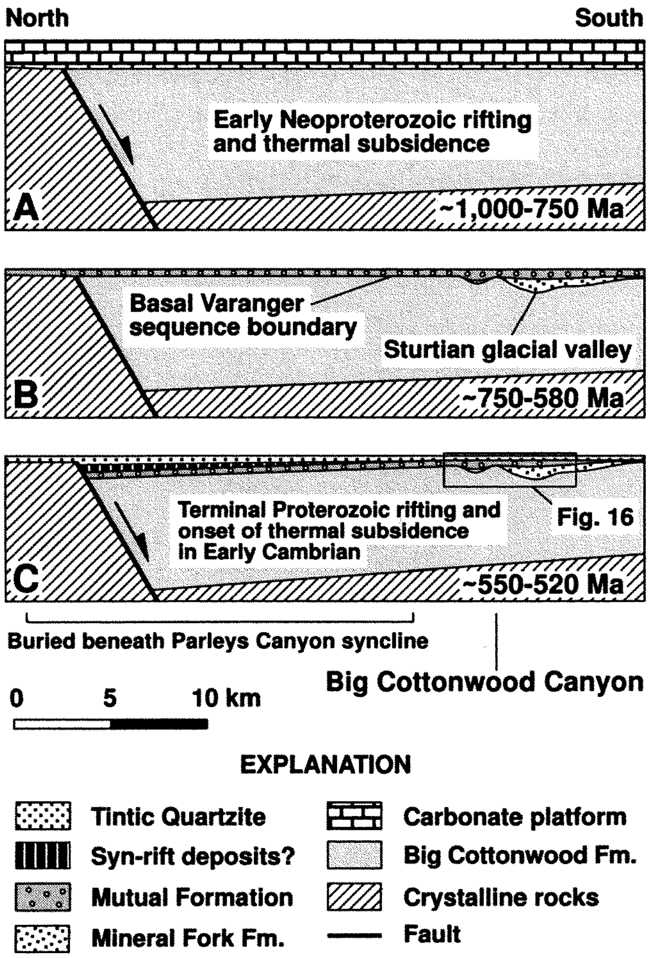

Figure 7. Neoproterozoic to Early Cambrian tectonic events interpreted in the central Wasatch Range near Big Cottonwood Canyon. A. Early Neoproterozoic rifting (Big Cottonwood Formation), and development of a carbonate platform during a phase of inter-rift thermal subsidence. B. Erosion of the platform during Sturtian glaciation (Mineral Fork Formation), and development of a basal Varanger sequence boundary (at this locality, at the base of the Mutual Formation). C. Renewed minor rifting and northward tilting of older rocks, and onset of post-rift thermal subsidence responsible for the early Paleozoic miogeoclinal succession. The Tintic Quartzite accumulated after rifting had ceased, but syn-rift deposits of terminal Proterozoic and Early Cambrian age may be present at depth beneath Parleys Canyon syncline. Note that the tilting shown (2°) is less than the 10° discordance observed locally beneath the Tintic Quartzite in Big Cottonwood Canyon. Vertical and horizontal scales are equal.

the presence of macrotidal rhythmites in the Big Cottonwood requires only a suitable connection with the open ocean. However, an inverted clast stratigraphy in the overlying Mineral Fork Formation (and in the Dutch Peak Formation of the Sheeprock Mountains) suggests that the siliciclastic rocks of the Big Cottonwood were originally overlain by a shallow-water carbonate platform, a succession of indeterminate thickness but which was progressively removed during glaciation (Fig. 7A and B; Blick, 1979; Christie-Blick, 1983b; Crittenden et al., 1983). A plausible interpretation of this inferred platform is that it developed during a phase of thermally driven subsidence after Big Cottonwood rifting had ended, implying a timing distinctly earlier than the phase of crustal extension recorded by the Sheeprock Group and correlative strata. Contact relations with the unconformably overlying Mineral Fork Formation also suggest that the Big Cottonwood Formation was lithified prior to Sturtian glaciation (Christie-Blick, 1983b; Christie-Blick and Link, 1988; Christie-Blick et al., 1989).

TINTIC QUARTZITE

Angular unconformities are unusual in the Neoproterozoic to Lower Cambrian succession of the western United States (Stewart, 1970; Crittenden et al., 1971). However, an example of a rift-onset unconformity, with an angular discordance of up to 10°, is present in the central Wasatch Range in the vicinity of Big Cottonwood Canyon (Fig. 7C; Christie-Blick and Levy, 1989a). The Tintic Quartzite cuts progressively downwards towards the south, resting successively on the Mutual Formation (from the mountain front to the vicinity of Big Cottonwood Canyon), the Mineral Fork Formation (at the head of Mineral Fork), and the Big Cottonwood Formation (south of Little Cottonwood Canyon). The degree of discordance decreases south of Mineral Fork but is difficult to estimate owing to the lack of useful markers in the Mineral Fork Formation. The implied direction of tilting is consistent with reactivation of the northern basin-bounding fault of the Big Cottonwood Formation.

The timing of this unconformity is uncertain owing to a hiatus of as much as 40 m.y. between deposition of the Mutual Formation (approximately 580 Ma) and the Tintic Quartzite (probably Lower Cambrian; < 543 Ma). I infer that tilting was coeval with the development in terminal Proterozoic time of the sequence boundary at the base of the Camelback Mountain Quartzite in southeastern Idaho and Geertsen Canyon Quartzite in northern Utah (Crittenden et al., 1971; Link et al., 1987; Levy and Christie-Blick, 1991a; Levy et al., 1994). Concordance of bedding between the Big Cottonwood and Mutual formations suggests that the Mutual predates the rifting event and is preserved within the graben because it was preferentially eroded from rift shoulders during the terminal Proterozoic (Fig. 7C). The preserved thickness of Mutual thus provides a minimum estimate of fault displacement (about 400 m). The thickness of the Tintic Quartzite on the north limb of Parleys Canyon syncline is comparable to that at Big Cottonwood Canyon (200–300 m; Crittenden et al., 1952; Lochman-Balk, 1976), in spite of a location above the footwall of the inferred basin-bounding fault. Two possible interpretations are: 1) the Tintic Quartzite is entirely post-rift, and syn-rift deposits are not preserved at this site far inboard of the main Cambrian depocenter; or 2) the Tintic sequence is partly syn-rift, but expanded sections are present in this area only in the subsurface in Parleys Canyon syncline (Fig. 7C).

DESCRIPTION OF TRAVERSES

In the following guide, each day is arranged in the form of a traverse, drawing attention to specific aspects of the geology at various locations or sections, but without a formal series of "stops." Elevations are given in metres (and feet), and traverse distances in kilometres (and miles). Distances along highways are quoted in miles only, as this is the conventional unit for vehicles and road maps. Stops for the first part of Day 3 in Big Cottonwood Canyon are described in Ehlers et al. (1997; their stops 1 to 5).

DAY 1: FRIDAY, 17 OCTOBER, 1997
Sheeprock Mountains (Harker Canyon to Dutch Peak)

Overview. Day 1 is devoted to a traverse at the northern flank of the Sheeprock Mountains (Figs. 1 and 3), with a focus on the interval from the base of the glacial-marine Dutch Peak Formation, through the Kelley Canyon Formation (basinal siltstone), to the basal part of the Caddy Canyon Quartzite (braided-fluvial pebbly quartzite and siltstone). Themes are evidence for Sturtian glaciation (~750 Ma), glacial-marine sedimentology and sequence stratigraphy, and Neoproterozoic rifting (Sheeprock Group).

Location of Traverse. Take I-80 west from Salt Lake City to exit 99, and then follow State Highway 36 south through Tooele to Vernon (45 miles from exit 99). Turn right (south) 0.7 mile beyond the gas station onto the graded gravel road towards the Sheeprock Mountains. Continue for 5.0 miles to Benmore, and turn right (west) at the T-junction. After 2.0 miles, turn left (south). Continue for a further 2.7 miles into Harker Canyon, and park at the sharp right-hand bend in the track (Figs. 3 and 8; Dutch Peak 7 1/2′ topographic map; 39° 58′ 46″ N., 112° 28′ 48″ W.). A vehicle with good ground clearance is needed for the final segment into Harker Canyon.

Traverse. Proceed up the ridge located immediately to the west of the parking place (Fig. 8). The traverse leads down section, from the poorly exposed lower Caddy Canyon Quartzite, through the Kelley Canyon Formation and the greywacke member of the Dutch Peak Formation. Continue east along the ridge towards Dutch Peak, and across a fault into the lower part of the diamictite member of the Dutch Peak Formation. The underlying conglomerate and

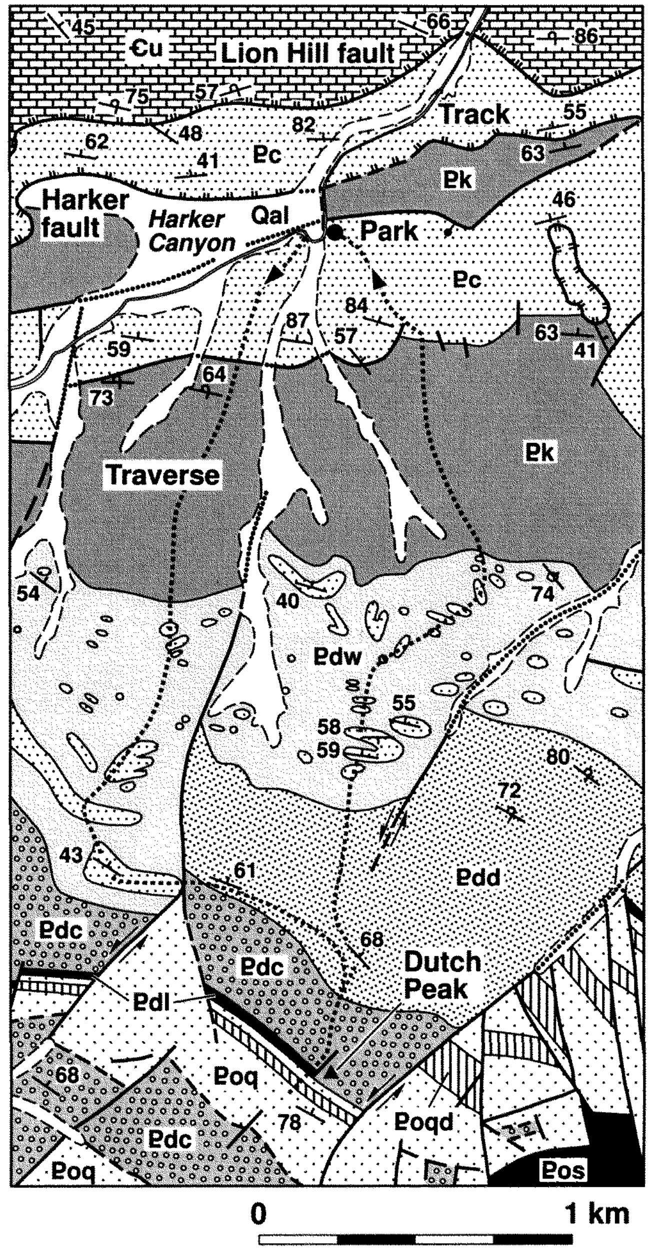

Figure 8. Geological map of the Harker Canyon-Dutch Peak locality, Sheeprock Mountains, showing the route of traverse for Day 1 (from Blick, 1979; Christie-Blick, 1982, 1983a). See Figure 3 for location and standard geological symbols. Abbreviations for map units as in captions for Figures 5 and 6. Other units: Ꝑoqd, diabase in the upper (quartzite) member of the Otts Canyon Formation; €u, undifferentiated Cambrian carbonate rocks; Qal, undifferentiated Quaternary alluvium. Stippled areas in the greywacke member of the Dutch Peak Formation (Ꝑdw) are quartzite bodies.

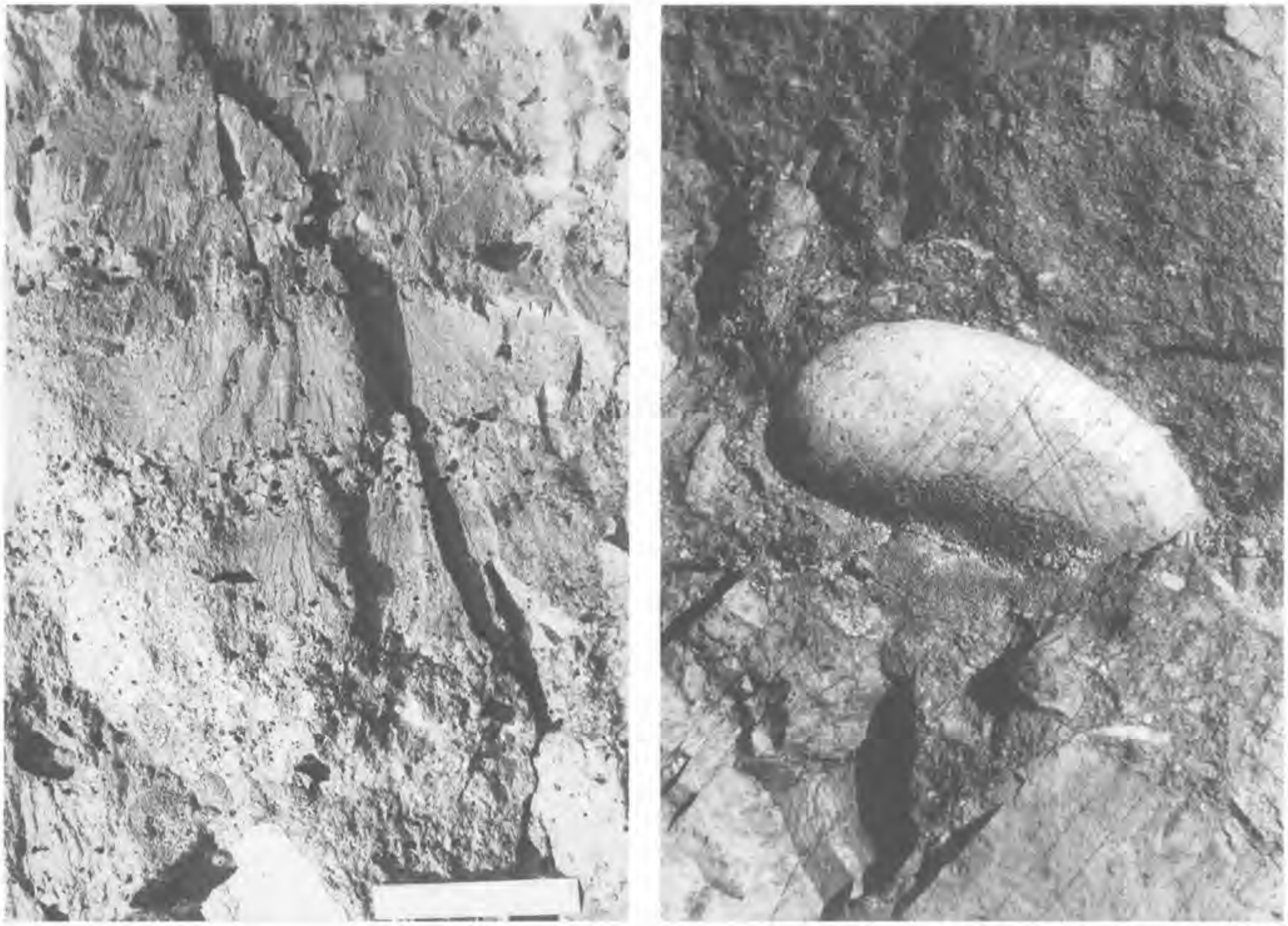

Figure 9. Stratified diamictite and conglomerate (A) and striated felsic volcanic cobble (B) in the conglomerate member of the Dutch Peak Formation in the vicinity of Dutch Peak, Sheeprock Mountains. The scale in A is 15 cm. The striated clast in B is 11 cm long.

"laminite" members are well exposed in the vicinity of Dutch Peak, together with the contact with the Otts Canyon Formation. The return journey to the vehicles allows us to walk through a complete section of Dutch Peak Formation, at its type location and in correct stratigraphic order. At the end of the day, we will briefly examine outcrops of the lower Caddy Canyon Quartzite immediately southeast of the parking place. The round trip to Dutch Peak (elevation, 2732 m or 8964 ft) is 7 km (4.4 miles), with an elevation change of about 650 m (2150 ft).

The Dutch Peak Formation is representative of glaciogenic deposits of Sturtian age (~750 Ma) that are widely exposed in the Cordillera of western North America (Stewart, 1972; Blick, 1979; Christie-Blick, 1982; Crittenden et al., 1983; Eisbacher, 1985; Link et al., 1994). It consists of as much as 1,750 m of crudely bedded to well-stratified diamictite, conglomerate, greywacke, sandstone, quartzite and siltstone (Fig. 9A). The matrix of the diamictite is typically olive green to grey, chloritic, moderately phyllitic and mineralogically, chemically and texturally heterogeneous at virtually all scales. Major clast types are dolomite, granitic rocks and quartzite (including a rare but distinctive green fuchsite quartzite). Subordinate types include various igneous, metamorphic and sedimentary rocks. Twenty-nine clast counts in the Sheeprock area indicate pronounced lateral and stratigraphic variations in the proportions of clast types over short distances (Blick, 1979), but some gross generalizations can be made. Dolomite clasts are most common in the diamictite member of the Dutch Peak Formation; quartzite clasts are most abundant in the greywacke member; and igneous and metamorphic clasts predominate in the conglomerate member. Representative clast counts for the Dutch Peak section are shown in Figure 10. Clasts are predominantly 2–5 cm in diameter, although rarely as large as 3 m. The size of clasts decreases stratigraphically upwards from boulder con-

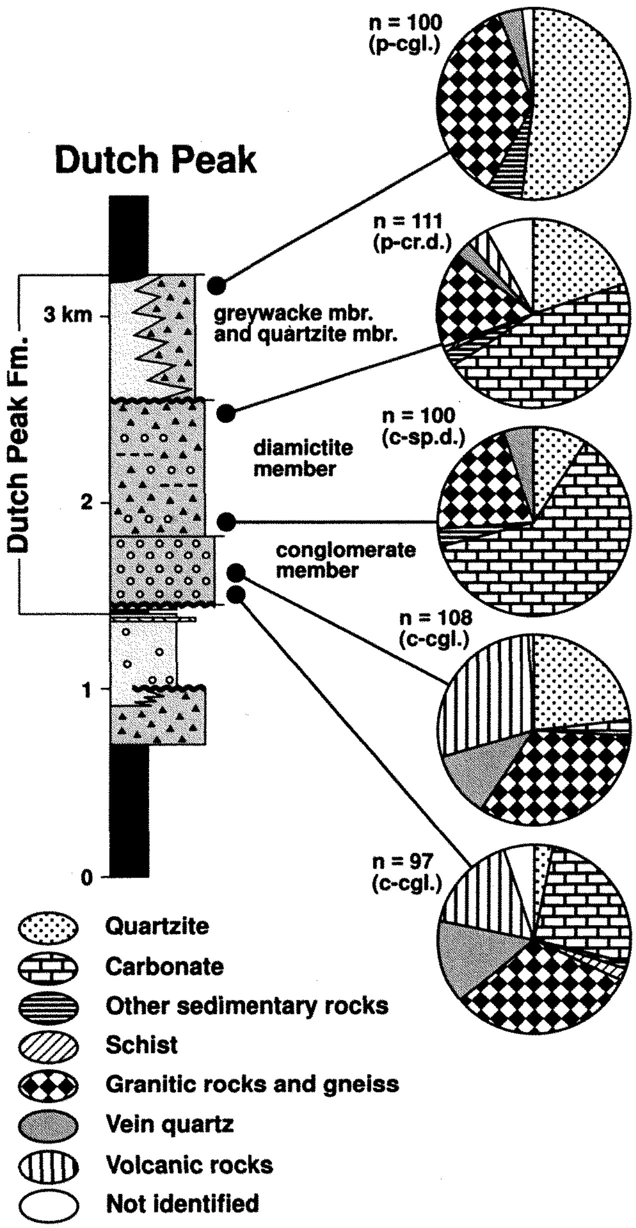

Figure 10. Clast counts (pie diagrams) for conglomerate and diamictite in the Dutch Peak section of the Dutch Peak Formation (localities 4 to 8 in Table 7 and Plate 7 of Blick, 1979). Minimum clast size counted, 1 cm. Abbreviations for dominant clast size and concentration: p, pebbles; c, cobbles; cgl., conglomerate; cr.d and sp.d, diamictite crowded with clasts and with relatively sparse clasts, respectively.

glomerate (conglomerate member) to sparsely pebbly diamictite (greywacke member). It has been possible to measure only a few paleocurrents: these are directed towards the southwest, approximately transverse to the direction of stratigraphic thickening in the upper plate of the Pole Canyon thrust.

Evidence supporting a glacial origin for these rocks includes the presence throughout the Cordillera of thick successions of similar diamictite, striated and faceted clasts, dropstones and isolated sediment pods in laminated fine-grained sedimentary rocks (inferred to have been ice-rafted), and beneath the Mineral Fork Formation of the central Wasatch Range, a grooved pavement overlain by diamictite (Christie-Blick, 1983b; Link et al., 1994). The Dutch Peak Formation contains sporadic examples of dropstones, especially in finely laminated siltstone and very fine-grained sandstone at the base of the formation ("laminite" member). Examples can be observed also in fine- to very fine-grained laminated quartzite in the uppermost part of the Otts Canyon Formation at Dutch Peak (Fig. 11A). A single striated cobble of felsic volcanic rock is present near the base of the conglomerate member, approximately 20 m south-southeast of the top of Dutch Peak (Fig. 9B). (Striated clasts are very rare in the Sheeprock Mountains. Please leave this one in place for others to examine!) The recent discovery, in Harker Canyon, of a clast of quartzite with shock lamellae led Oberbeck et al. (1994) to suggest that the Dutch Peak Formation might not be glaciogenic after all, but an impact ejecta blanket. Given the overall sedimentological and stratigraphic context of the rocks, a more plausible interpretation is that an indeterminate quantity of impact material was simply eroded by the ice sheet.

A basinal glacial-marine setting is inferred for much or all of the Sheeprock Group, although in places the ice sheet appears to have been grounded below sea level. As outlined above, rocks of the conglomerate member of the Dutch Peak Formation overlie a deep-water sequence boundary, and are tentatively interpreted as an ice-proximal subaqueous fan that developed during glacial advance. The remainder of the Dutch Peak Formation accumulated during glacial retreat, although sedimentation was punctuated by a second sequence boundary that can be traced to near the contact between the diamictite and greywacke members. This contact is best examined on the traverse towards Dutch Peak, where the lower part of the greywacke member contains several prominent lenses of quartzite, here interpreted as ice-proximal glacial-marine outwash. The diamictite member appears to pinch out (perhaps by onlap) towards the northwest in the same area, although stratigraphic details are obscured by faulting.

The Dutch Peak Formation and the underlying Otts Canyon Formation are characterized by some marked changes in thickness and facies from Harker Canyon to the western side of the range (described above). These changes are difficult to demonstrate in a single day on the outcrop. However, some appreciation can be gained from the vantage of Dutch Peak: the Dutch Peak Formation thins by a factor of three in a distance that is approximately

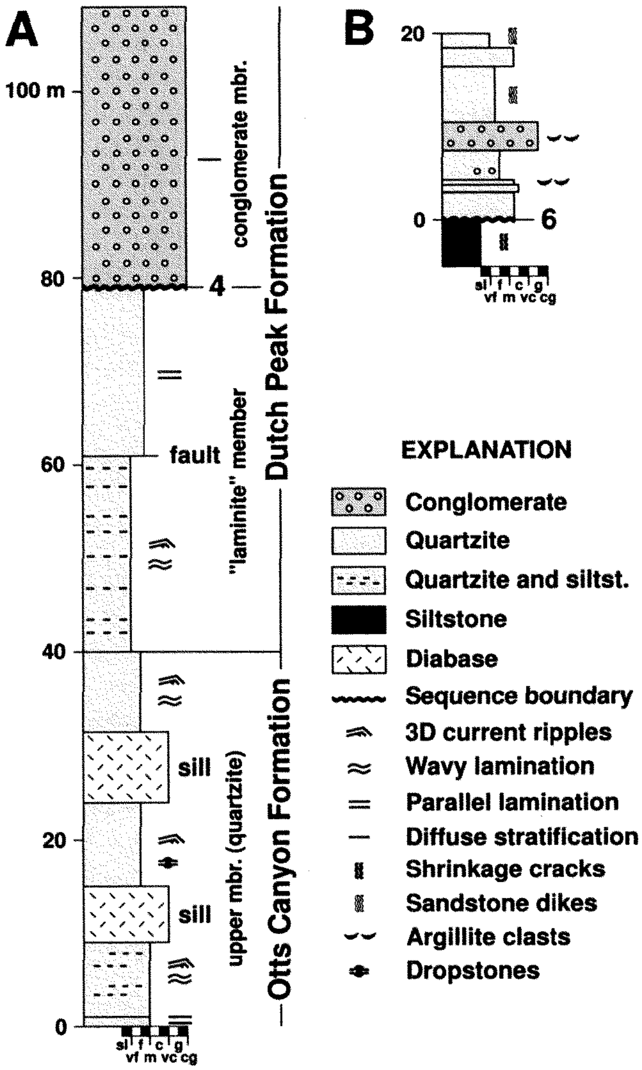

Figure 11. Simplified stratigraphic sections through the transition between the Otts Canyon and Dutch Peak formations at Dutch Peak (A) and at the contact between the Kelley Canyon Formation and Caddy Canyon Quartzite (sequence boundary 6; B) at the end of the Day 1 traverse (see Fig. 8). Column width refers to the Wentworth grain-size scale (sl, siltstone; vf, f, m, c and vc, sandstone range; g and cg, granule and coarser conglomerate).

twice the distance to the head of Harker Canyon on the skyline to the northwest. In that same area it is possible to recognize northwestward thinning of the greywacke member, and its lateral change in facies to quartzite in the same direction. The overall pattern is ascribed to differential subsidence adjacent to extensional growth faults, although the faults themselves are not exposed in available outcrop. Mafic sills, contemporaneous with deposition, and perhaps related to crustal extension, are exposed near the top of the Otts Canyon Formation at Dutch Peak (Fig. 11A).

The transition from the Dutch Peak Formation to laminated siltstone of the Kelley Canyon Formation is thought to be due to a reduction in the supply of terrigenous sediment, perhaps combined with a glacial-eustatic rise. Similar rocks overlie Sturtian-age glacial deposits throughout the Cordillera (and in fact, globally). In many basins, the transition is marked by a deep-water laminated micrite or dolomicrite (a so-called "cap carbonate"). Examples are pres- ent in Utah at Huntsville and on Antelope Island in the Great Salt Lake (Fig. 1), but not in the Sheeprock Mountains.

In many places, the contact between the Kelley Canyon Formation and Caddy Canyon Quartzite is transitional, representing gradual shoaling during progradation of nearshore sediments (Christie-Blick and Levy, 1989a; Levy and Christie-Blick, 1991a). A sharp contact in the northern part of the Sheeprock Mountains is interpreted to represent a locally developed sequence boundary (surface 6 in Figs. 2 and 11B). In Harker Canyon, the basal beds of the Caddy Canyon Quartzite consist of braided-fluvial pebbly quartzite and siltstone, including some impressive sandstone dikes. This outcrop foreshadows the second and third days of the excursion, in which similar facies (minus the dikes) will be found to be characteristic of incised valley fills within the upper part of the Brigham Group.

DAY 2: SATURDAY, 18 OCTOBER, 1997
Sheeprock Mountains (Horse Valley)

Overview. The focus of Day 2 is the sedimentology and sequence stratigraphy of the interval from the upper part of the Caddy Canyon Quartzite, through the Inkom Formation (braided-fluvial conglomerate and marine siltstone), to the basal part of the Mutual Formation (also braided-fluvial conglomerate). Themes are incised valleys and incised valley fills in a braid-deltaic setting, the origin and interpretation of Proterozoic sheet quartzites (braided fluvial versus shallow marine), and the expression of Varanger glaciation (~600 Ma) in non-glacial facies.

Location of Traverse. Proceed southward along State Highway 36 from the overnight stop in Tooele to Vernon (about 32 miles). As on Day 1, turn right (south) 0.7 mile beyond the gas station onto the graded gravel road towards the Sheeprock Mountains. Continue for 5.0 miles to Benmore, and then turn left (east) at the T-junction. Continue for 9.6 miles southward along Vernon Creek, turning right (southwest) off the graded gravel road 0.2 mile beyond the dry farm. Turn right at the next junction (0.2 mile), and continue along that track for an additional 1.0 mile. Turn right again (north), and continue for 1.0 mile to a gate in Horse Valley. After 1.1 mile, turn left (south) through another gate, and continue for 0.9 mile along a jeep track.

Park at the crest in the track, and continue on foot towards the south (Figs. 3 and 6; Dutch Peak 7 1/2′ topographic map; 39° 53′ 26″ N., 112° 23′ 59″ W.). The jeep track is over level country, but requires a vehicle with good ground clearance.

Traverse. Follow the track for 2 km (about 1 mile) towards the southwest, to the adit at the contact between the Caddy Canyon Quartzite and Inkom Formation (section f and surface 7 in Fig. 6A and C). All of the rocks are overturned and dip at about 25° towards the west. Trace the top of the quartzite southwards to section g, noting the facies of the quartzite, evidence for erosional truncation, and onlap of the basal Inkom conglomerate. Walk through section g (downhill), as far as the lower part of the Mutual Formation. Trace the upper siltstone unit of the Inkom Formation northward to section e. Compare the stratigraphy of section g with that at sections b and a (Fig. 6A). The upper Inkom siltstone is unusually thin at section b (4 m) owing to erosional relief at the base-Mutual sequence boundary (surface 9 in Fig. 6A and C). The underlying conglomerate unit is unusually thin at section a (3.5 m), and pinches out between sections e and c where the track crosses the outcrop. The round trip is about 5 km (3 miles) from the specified parking place, at an elevation of about 2,125 to 2,250 m (7,000 to 7,400 ft). It is possible to drive as far as section f (Fig. 6A) with a four-wheel-drive vehicle.

The contact at the base of the Inkom Formation is associated with local erosional relief of several tens of metres and a very pronounced change in facies (Figs. 12 and 13A; Levy et al., 1994). The upper part of the Caddy Canyon Quartzite is composed of vitreous white to grey coarse- to fine-grained quartzite with indistinct trough cross-stratification. The (stratigraphically) overlying Inkom Formation consists at the base of greyish-red pebbly quartzite and pebble conglomerate with outsized clasts of liver-colored and bluish-grey argillite (< 3 m). The rocks are broadly channelized, with diffuse stratification picked out primarily by alignment and variations in the size and abundance of clasts. One large clast of laminated argillite contains a layer of pebble conglomerate, itself containing clasts of argillite. Evidently, fine-grained sediments accumulated in the valleys during times of low discharge, only to be reamed out during times of flood. Rounded pebbles consist primarily of vein quartz, but with approximately 30% quartzite resembling the underlying Caddy Canyon Quartzite (Fig. 6D). This, and the absence of deformation at the valley walls, suggests that the Caddy Canyon Quartzite was consolidated or lithified at the time of valley incision. Similar although less coarse-grained conglomerate is present at incised valleys in the upper part of the Inkom Formation and at the base of the Mutual Formation, in both cases cutting into predominantly fine-grained deposits (surfaces 8 and 9 in Figs. 6 and 12; Fig. 13B).

The abundance of stratification and channels in conglomerate and pebbly quartzite, and the absence of turbidites in associated siltstone, indicates that the coarse-grained deposits accumulated in a fluvial to shallow marine environment, and primarily as a result of river floods. (We cannot exclude a debris-flow origin for some of the beds.) The depth of the valleys (30–160 m), and the fact that those within the Inkom and at the base of the Mutual cut into marine siltstone, is consistent with large-amplitude sea-level change during a time of glaciation (elsewhere). These valleys are also not consistent with the standard model in which valley incision is attributed to headward erosion from a break in the depositional profile near the shoreline (see Posamentier et al., 1988; Levy et al., 1994; Christie-Blick and Driscoll, 1995). Evidently, the valleys within the Inkom Formation and at the base of the Mutual Formation extended well seaward of the highstand shoreline.

The Caddy Canyon Quartzite is representative of huge thicknesses of Proterozoic and Lower Cambrian rocks that are commonly interpreted as having accumulated in a shallow marine environment. Some of these quartzite units display good evidence for tidal activity (e.g., Big Cottonwood Formation) or for deposition on wave-dominated shelves (e.g., Lower Cambrian Campito Formation, California; Mount, 1982). However, thick successions that extend areally for hundreds of kilometres and consist virtually entirely of quartzite may be better interpreted as braided fluvial deposits. This is especially the case when paleocurrent trends appear to be transverse to regional shorelines (as in the Brigham Group; Levy and Christie-Blick, 1991a). Moreover, sequence stratigraphic studies show that the sheet-quartzite facies of the Caddy Canyon Quartzite seen at Horse Valley is paleogeographically intermediate between fluvial deposits to the east and northeast and marine deposits to the west (Levy and Christie-Blick, 1991a). The facies of such quartzites tend to change gradually, both vertically and laterally, from channelized, with thinning- and fining-upward successions (fluvial), to thin- and tabular-bedded, with interstratified siltstone and (in many Cambrian examples) marine trace fossils (e.g., Lower Cambrian Wood Canyon Formation, California; Diehl, 1976). The model that best fits the intermediate facies is that of a braid delta in a ramp setting, in which distal braided rivers are characterized by broad shallow channels, and the shallow marine environment is strongly influenced by fluvial input during times of flood.

DAY 3: SUNDAY, 19 OCTOBER, 1997
Central Wasatch Range (Big Cottonwood Canyon)

Overview. The final day of the excursion is divided into two parts. The first part consists of stops 1 to 5 of Ehlers

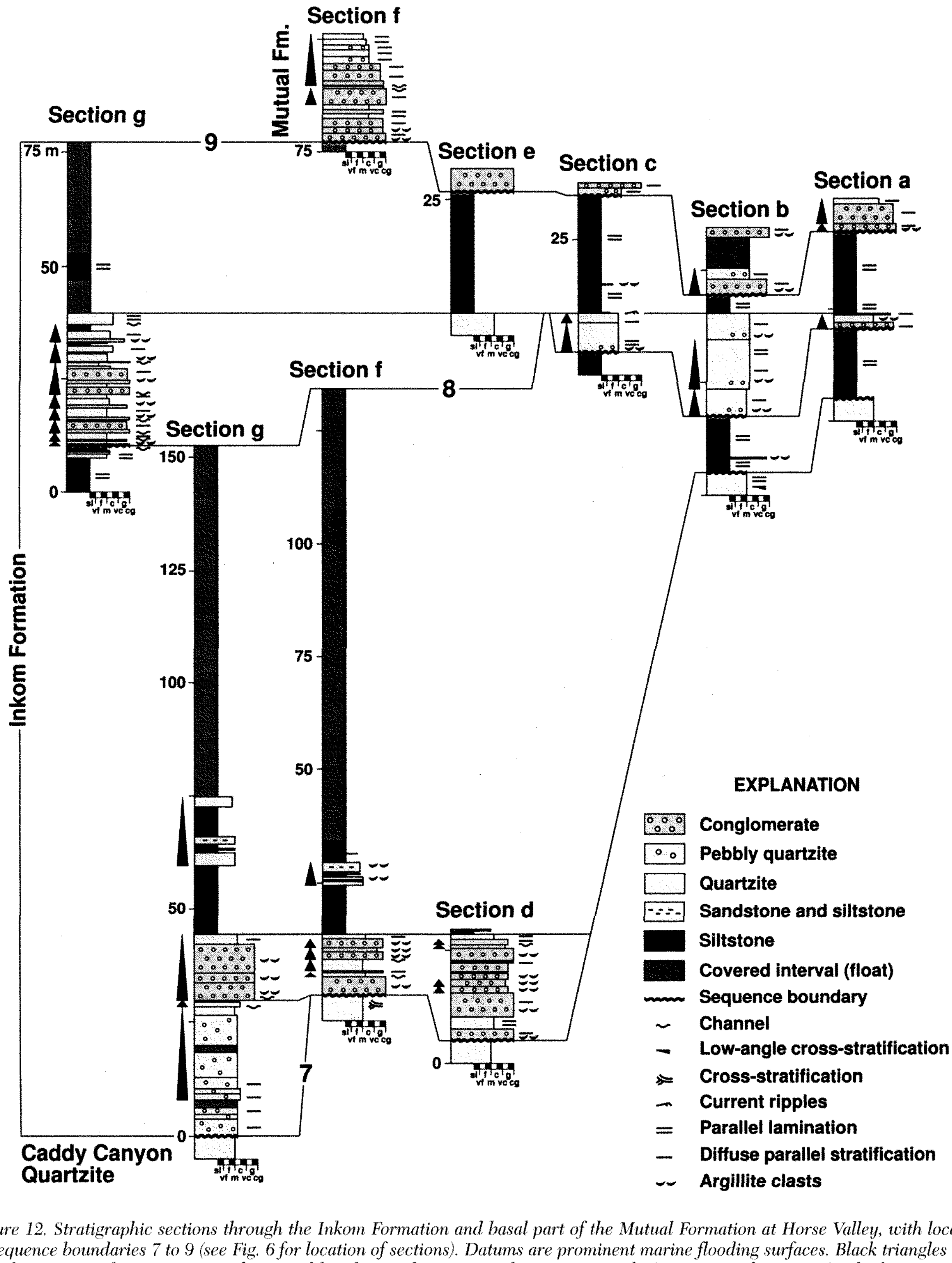

Figure 12. Stratigraphic sections through the Inkom Formation and basal part of the Mutual Formation at Horse Valley, with location of sequence boundaries 7 to 9 (see Fig. 6 for location of sections). Datums are prominent marine flooding surfaces. Black triangles indicate fining-upward successions. Column width refers to the Wentworth grain-size scale (see caption for Fig. 11). The lower parts of sections f and g were measured with M. Levy (see Levy and Christie-Blick, 1991a; Levy et al., 1994). The other sections were measured by the author in 1996.

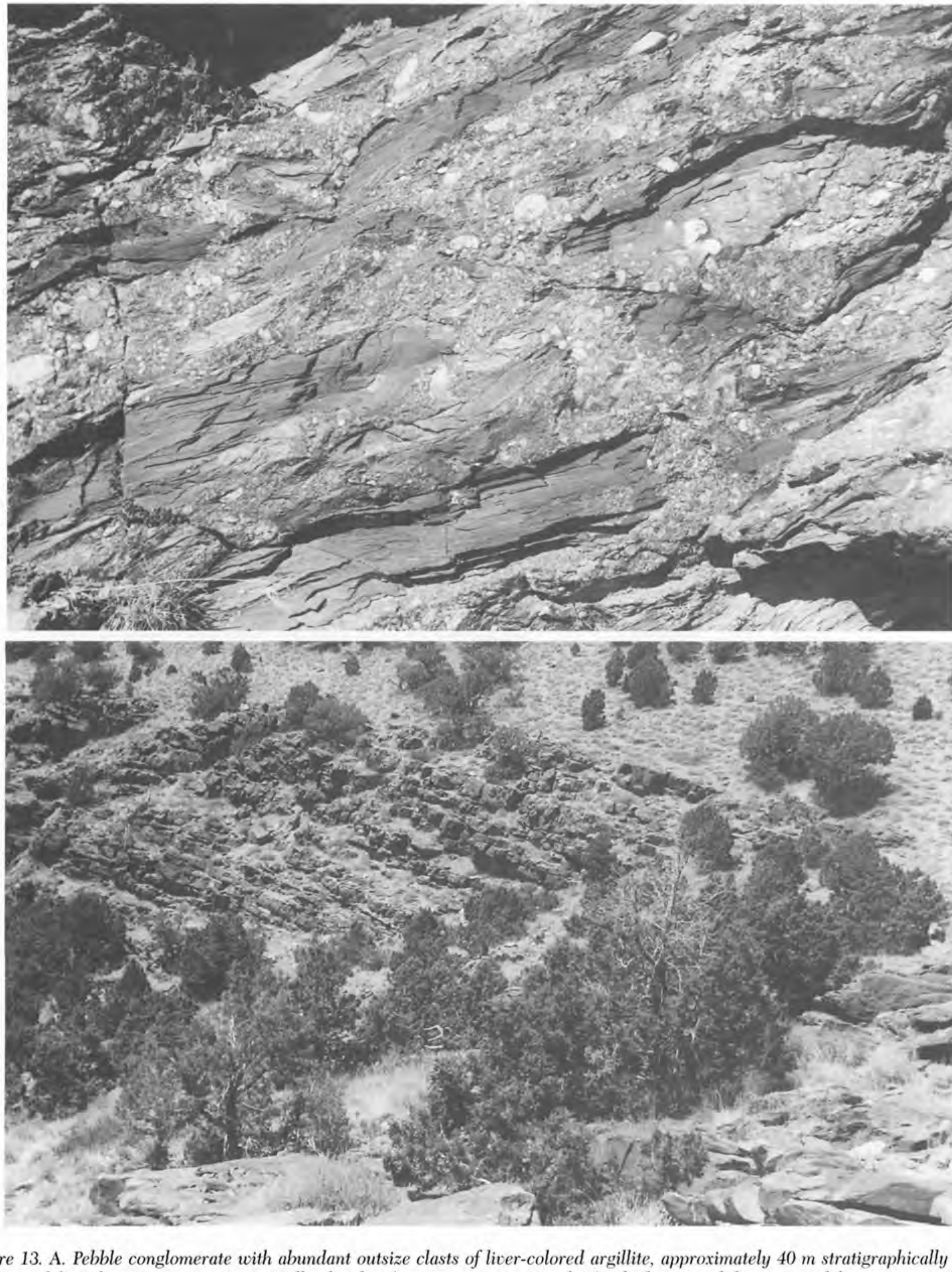

Figure 13. A. Pebble conglomerate with abundant outsize clasts of liver-colored argillite, approximately 40 m stratigraphically above the base of the Inkom Formation, Horse Valley locality (section g in Figs. 6 and 12). The horizontal dimension of the outcrop is about 1 m. B. Parallel-stratified quartzite and pebble conglomerate above sequence boundary 8, in the middle part of the Inkom Formation (near section g in Figs. 6 and 12). The rocks in both photographs are overturned.

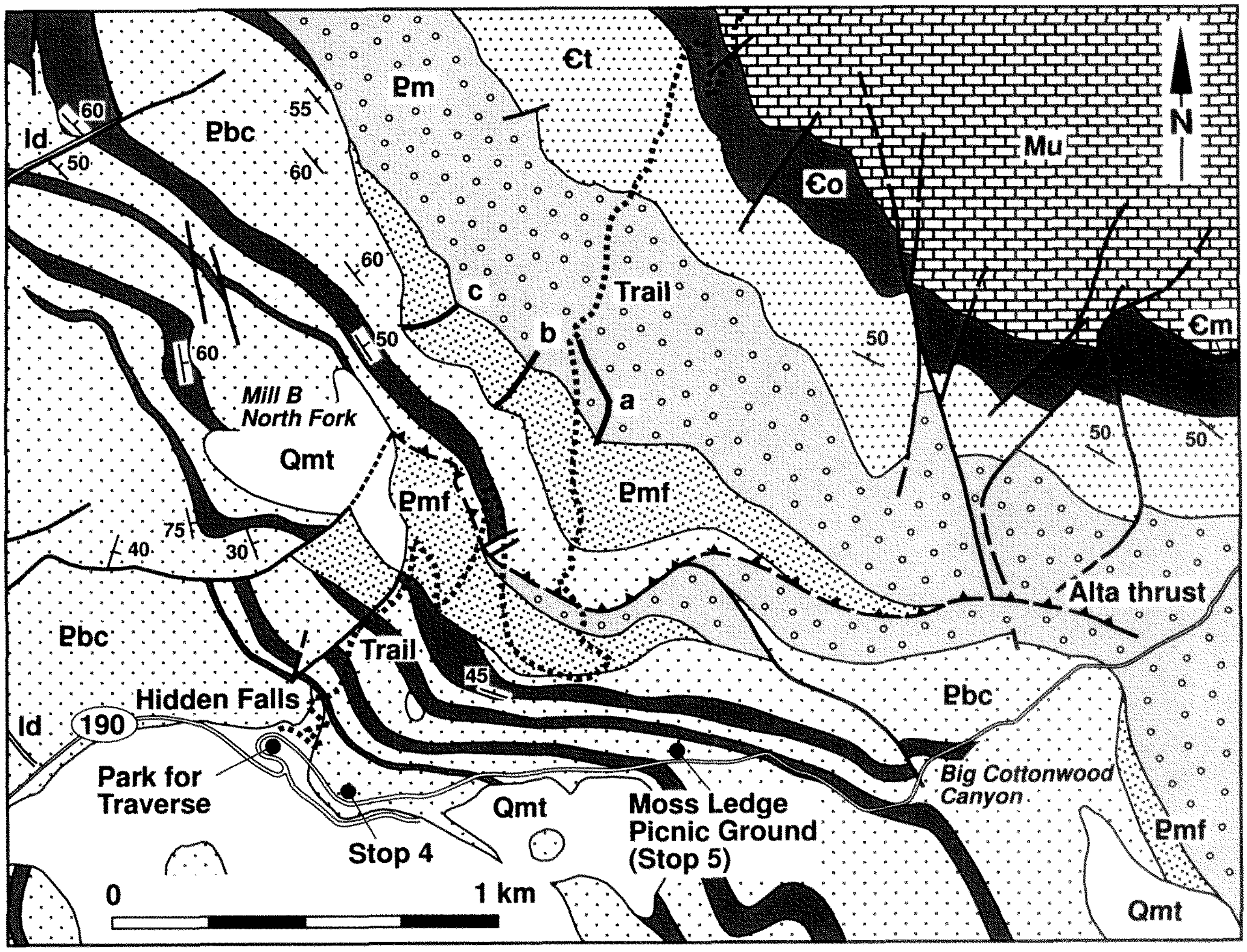

Figure 14. Geological map of Mill B North Fork of Big Cottonwood Canyon, central Wasatch Range (from Crittenden, 1965a), showing the trail from Hidden Falls and stratigraphic sections a to c for Day 3 (see Fig. 15 below). Also shown are field excursion stops 4 and 5 of Ehlers et al. (1997). See Figure 1 for location. Standard geological symbols are used. Stratigraphic units are as follows. Proterozoic: P̲bc, Big Cottonwood Formation; P̲mf, Mineral Fork Formation; P̲m, Mutual Formation. Cambrian: €t, Tintic Quartzite; €o, Ophir Formation; €m, Maxfield Limestone. Other units: Mu, undifferentiated Mississippian rocks; ld, lamprophyric dikes (Cretaceous or Tertiary); Qmt, undifferentiated Quaternary glacial moraine and talus.

et al. (1997), with a focus on quartzite-siltstone tidal rhythmites of the Mesoproterozoic to Neoproterozoic Big Cottonwood Formation. The second part involves a traverse at Mill B North Fork of Big Cottonwood Canyon from the upper part of the Big Cottonwood Formation through the glacial-marine Mineral Fork Formation to the lower part of the Mutual Formation (mostly braided-fluvial conglomerate and quartzite). Themes for the latter are Sturtian glacial-marine sedimentation in glacially incised valleys, correlation of the upper Caddy Canyon and base-Mutual sequence boundaries into the central Wasatch Range, and evidence for Mesoproterozoic and Neoproterozoic rifting (Big Cottonwood Formation and the angular unconformity beneath the Tintic Quartzite).

Location of Stops and Traverse. Take I-80 east from the overnight stop in Tooele towards Salt Lake City. After passing Salt Lake City International Airport, follow I-215 south and east towards the front of the Wasatch Range. At exit 7, take State Highway 190 south to Big Cottonwood Canyon. We meet participants of the one-day Big Cottonwood excursion in the parking area at the mouth of the canyon (Fig. 1; and Fig. 2, Stop 1 of Ehlers et al., 1997). All mileages are measured along Highway 190 (formerly 152) east of this point. The afternoon traverse at Mill B North

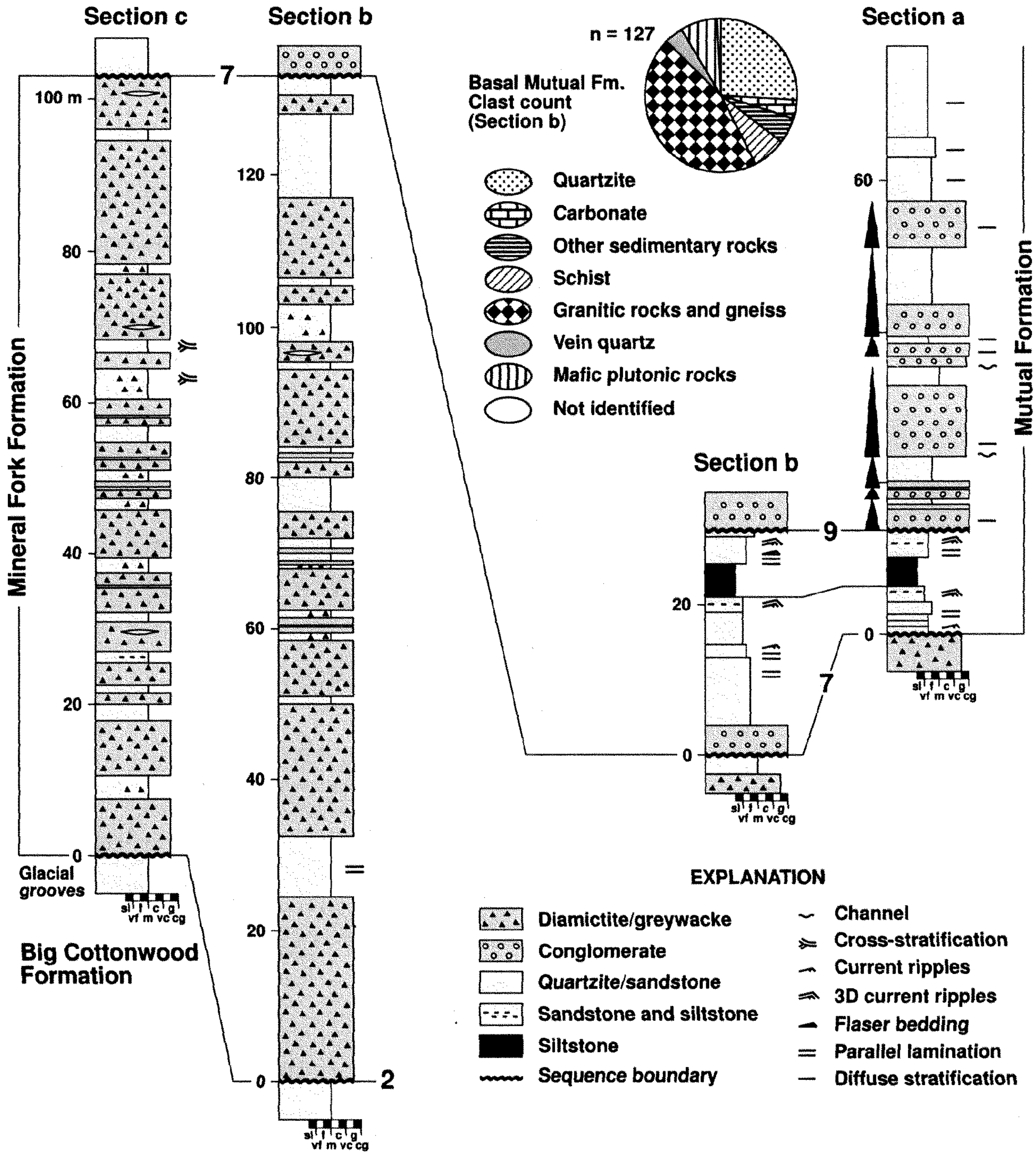

Figure 15. Stratigraphic sections through the Mineral Fork Formation and lower part of the Mutual Formation at Mill B North Fork of Big Cottonwood Canyon (from Blick, 1979; Levy and Christie-Blick, 1991a; Levy et al., 1994; and unpublished data; see Fig. 14 for location of sections). Datums are sequence boundaries 7 and 9. Black triangles indicate fining-upward successions. Column width refers to the Wentworth grain-size scale (see caption for Fig. 11). Grain sizes for Mineral Fork Formation are approximate only. Clast count (pie diagram) for basal conglomerate of the Mutual Formation. Minimum clast size counted, 1 cm.

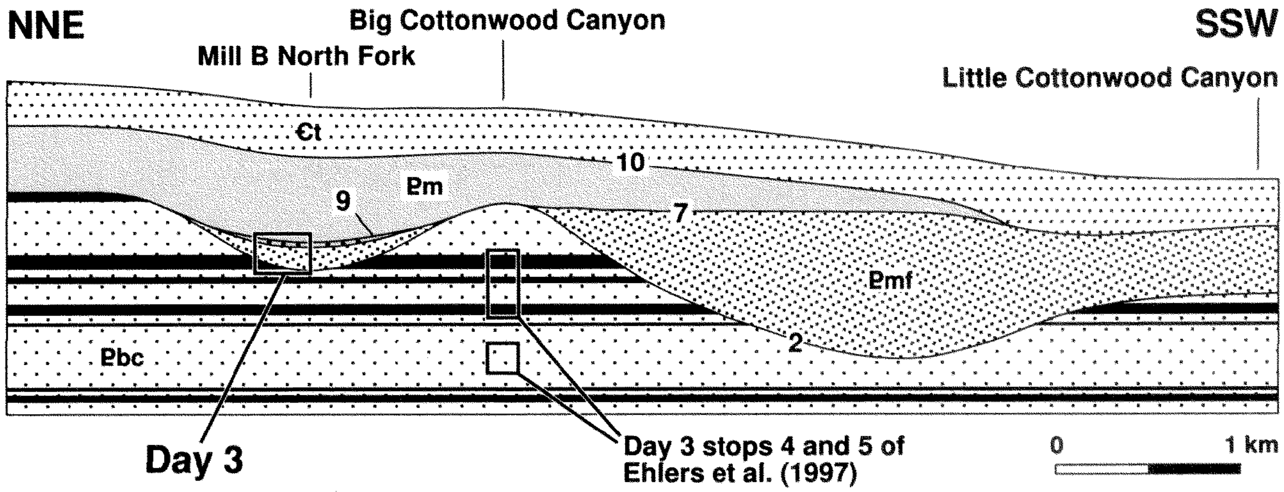

Figure 16. Palinspastic cross section of Sturtian-age glacial valleys filled by the glacial-marine Mineral Fork Formation (Ꝑmf) in the vicinity of Big and Little Cottonwood Canyons (based on mapping by Crittenden ,1965a, 1965d, and modified from Christie-Blick, 1983b). Bedding in the Big Cottonwood Formation is taken as paleohorizontal. Note that the northern valley, to be examined on Day 3, is repeated by the Alta thrust (see Fig. 14). The thickness of Mineral Fork Formation shown is appropriate for the upper plate of the thrust (see Fig. 15), but the deeper valley profile depicted here is preserved in the lower plate. Quartzite and siltstone units in the Big Cottonwood Formation (Ꝑbc) are indicated by stipple and black fills, respectively, with thicknesses taken from the mapping of Crittenden (1965a, 1965d) in Big Cottonwood Canyon. These thickness estimates differ slightly from measurements at road level by Ehlers et al. (1997). Ꞓt, is Tintic Quartzite (Cambrian). Vertical and horizontal scales are equal.

Fork begins at the S curve (Hidden Falls) at mile 4.6 (Fig. 14; Mount Aire 7 1/2' topographic map; 40° 38' 03" N., 111° 43' 24" W.). See also the Draper, Sugar House and Dromedary Peak 7 1/2' topographic maps, and corresponding geological maps by Crittenden (1965a, 1965b, 1965c, 1965d). The latter have been compiled into a single sheet at 1:24,000 scale by James (1979).

Stop 1: Overview of the Wasatch Range (mile 0)
Stop 2: Remnants of an Ancient Sea Sign (mile 2.3)
Stop 3: Storm Mountain Quartzite Sign (mile 2.9)
Stop 4: S Curve Channeled Quartzite (mile 4.7)
Stop 5: Moss Ledge Picnic Area (mile 5.1)

Return to the S curve (Hidden Falls), and park in the designated area at mile 4.6 (Fig. 14).

Traverse. The main outcrops in Mill B North Fork are located about 425 m (1,400 ft) above road level, at an elevation of 2,250 to 2,375 m (7,400 to 7,800 ft), and are reached on the north side of Big Cottonwood Canyon via a steep switch-back trail beyond Hidden Falls (Fig. 14; Crittenden, 1965a). The Mineral Fork Formation is exposed in two panels owing to repetition by the Alta thrust. The lower panel is poorly exposed, and in the time available we will focus on the outcrops of the Mineral Fork Formation and lower part of the Mutual Formation above the Alta thrust. Most accessible are sections of Mutual Formation on both sides of the trail (sections a and b in Fig. 15). If time is available, we may also examine one or both of the sections through the Mineral Fork Formation (sections b and c in Fig. 15). Glacial grooves of Proterozoic age preserved on top of the Big Cottonwood Formation at section c.

The Mineral Fork Formation consists of up to 800 m of diamictite, siltstone, sandstone and conglomerate (Crittenden et al., 1952; Ojakangas and Matsch, 1980; Christie-Blick, 1983b; Christie-Blick and Link, 1988; Christie-Blick et al., 1989; Link et al., 1994). North of Big Cottonwood Canyon, the rocks are comparatively fine-grained, consisting predominantly of sparsely pebbly diamictite (greywacke), with lenses of medium- to fine-grained sandstone (in places cross-stratified) and laminated siltstone.

The rocks occupy two west-northwest trending paleovalleys eroded into the top of the Big Cottonwood Formation (Fig. 16). The southern valley is about 900 m deep and more than 7 km wide. The northern valley is between 200 and 425 m deep (two sections) and about 2 km wide, and the one to be examined on this excursion. Note that the deeper section through the northern valley is in the lower plate of the Alta thrust. Assuming tectonic transport towards the east, as is implied by eastward-climbing footwall and hanging-wall ramps, then the valley floor was inclined towards the east-southeast, a direction that is opposite to the azimuth of paleocurrents within the Mineral Fork Formation (approx. 290°; Christie-Blick, 1983b; Christie-Blick et al., 1989).

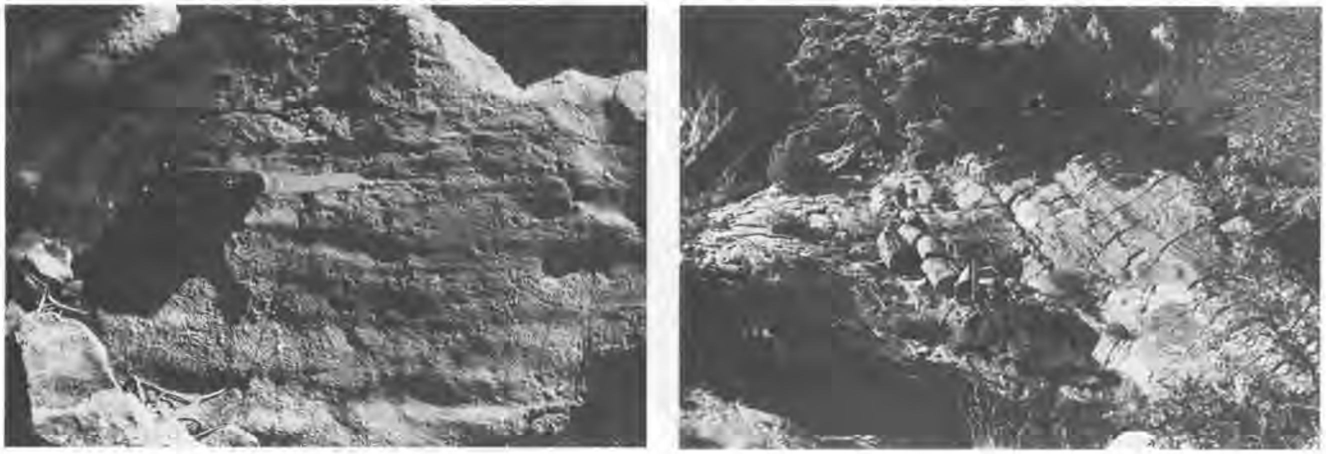

Figure 17. A. Grooves of inferred Proterozoic glacial origin at the contact between the Big Cottonwood Formation and Mineral Fork Formation, Mill B North Fork of Big Cottonwood Canyon (section c in Figs. 14 and 15). B. Linear ridges parallel to the grooves are interpreted as roches moutonneés, with glacial flow towards the northwest (308° after correction for tectonic tilt).

The contact between the Mineral Fork Formation and underlying Big Cottonwood Formation is knife-sharp, with no sign whatsoever of soft-sediment deformation in spite of local paleogradients as great as 40°. This, the scale of the valleys, and the presence of rounded clasts of Big Cottonwood quartzite in diamictite indicates that the Big Cottonwood Formation was lithified at the time of glaciation, consistent with an age significantly older than the Mineral Fork Formation. At section c (Fig. 15), the contact is characterized by erosional ridges up to 2 m high (interpreted as roches moutonnées), with grooves as much as 1 m long oriented approximately parallel to the ridges (Fig. 17; trend corrected for tectonic tilt, 308°). The grooves are also approximately parallel to the traces of cross-stratification in the Big Cottonwood Formation, but unlike the cross-stratification they pass through bedding planes. The grooves are oriented transverse to the local flow direction of Pleistocene glaciers. I infer that the surface was glacially striated in Neoproterozoic time.

The contact between the Mineral Fork and Mutual formations is characterized by erosional relief of more than 15 m, and is overlain at section b by a 4-m-thick disorganized polymictic boulder conglomerate that appears to have been derived primarily from the underlying diamictite (Figs. 15 and 18A). The conglomerate is overlain in turn by a fining-upward succession, a few metres thick, of fine- to very fine-grained quartzite and siltstone with even parallel laminae and current ripples. Above this level, the rocks coarsen upwards over several metres to a second, rather thicker interval of conglomerate and quartzite (Figs. 15 and 18B). This upper conglomerate is crudely stratified, in places channelized and characterized by fining-upward successions up to a few metres thick. The coarse-grained rocks are interpreted as primarily fluvial (and possibly debris-flow) deposits corresponding with the upper Caddy Canyon and base-Mutual sequence boundaries of the miogeocline (surfaces 7 and 9 in Fig. 2). That is, the feather edge of the Inkom sequence is present in the lower part of the Mutual Formation at Mill B North Fork. The lithostratigraphic base of the Mutual (the upper Caddy Canyon sequence boundary) corresponds here with a hiatus of more than 100 m.y., and is thought to reflect a combination of late Sturtian glacial-isostatic uplift and Varanger-age glacial-eustatic drawdown (Fig. 7B).

The terminal Proterozoic rift-onset unconformity at the base of the Tintic Quartzite may be reached further along the trail approximately 600 m (2,000 ft.) from the base of the Mutual Formation. Angular relations with the underlying Big Cottonwood-Mutual succession are best seen at the mountain front north of Big Cottonwood Canyon (Fig. 19). The most favorable lighting for this view is late afternoon; in October we shall undoubtedly have run out of daylight before reaching the Salt Lake Valley.

ACKNOWLEDGMENTS

This field guide is a byproduct of research supported by the National Science Foundation, by the Donors of the Petroleum Research Fund, administered by the American Chemical Society, and by the Arthur D. Storke Memorial Fund of the Department of Earth and Environmental Sciences, Columbia University. I have benefitted over the years from discussions with numerous colleagues and students concerning both general geological issues and specific outcrops covered by this guide. In particular, I acknowledge the contributions of the late Max D. Crittenden, Jr. and Hal T. Morris (U.S. Geological Survey), who were responsible for primary geological mapping in the Wasatch Range

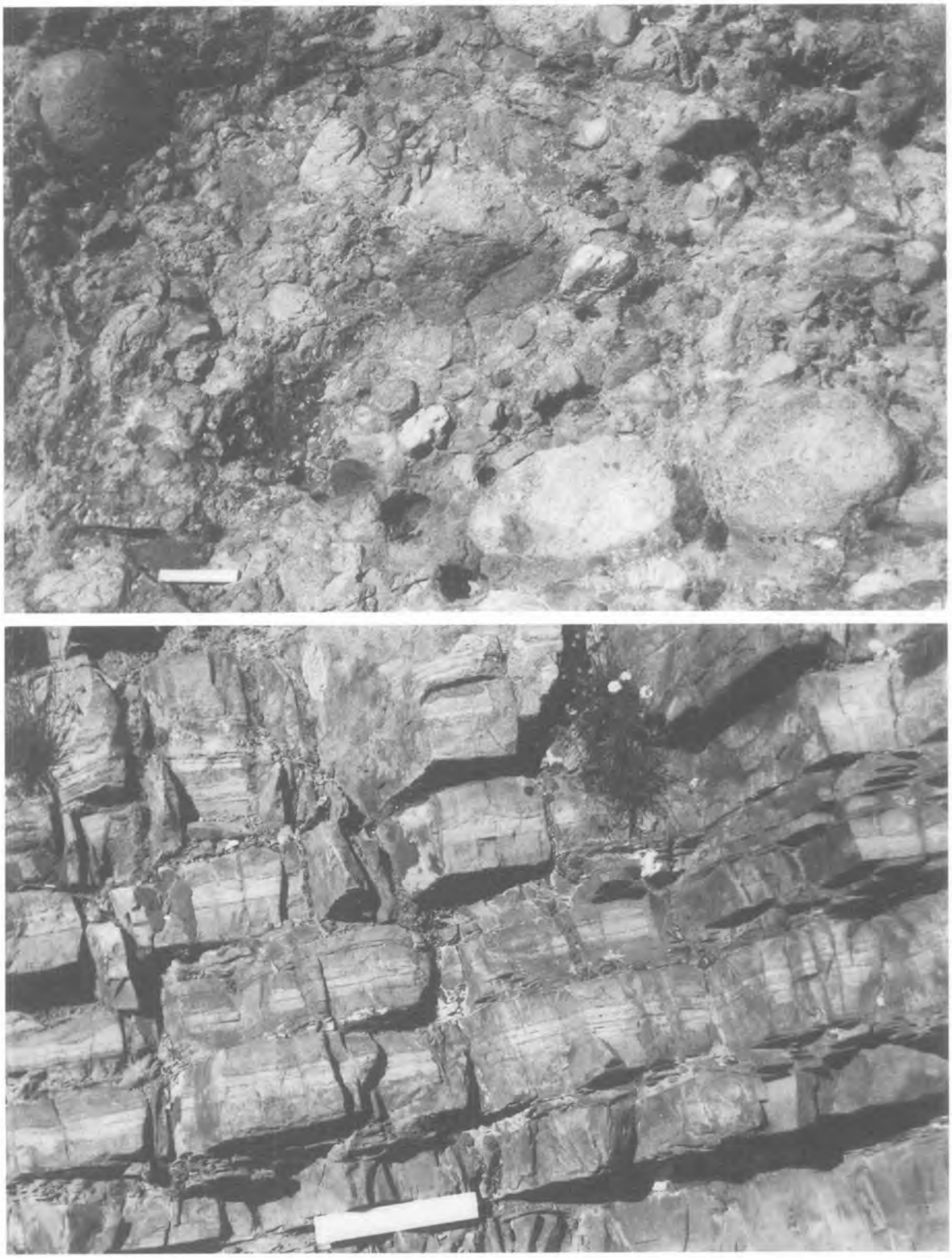

Figure 18. A. Disorganized boulder conglomerate at the base of the Mutual Formation, immediately above sequence boundary 7, Mill B North Fork (section b in Figs. 14 and 15). B. Fine- to very fine-grained quartzite and siltstone in the Mutual Formation, immediately below sequence boundary 9 (section a in Figs. 14 and 15). The quartzite beds are event layers, with subtle normal grading, and diffuse to well developed parallel lamination in lower parts of beds and 3D current ripples in the upper parts of beds. The scale in each photograph is 15 cm.

Figure 19. The mountain front of the Wasatch Range, immediately north of Big Cottonwood Canyon. The contact between the Big Cottonwood Formation and Mutual Formation (tree-filled gully in the center of the photograph) is concordant, in spite of a hiatus in sedimentation of up to several hundred million years. The Mutual Formation is overlain with angular unconformity, to the left, by less steeply dipping beds of the Tintic Quartzite (Cambrian). The base of the Tintic is interpreted as a rift-onset unconformity.

and southern Sheeprock Mountains, respectively. Sequence stratigraphic studies in the Proterozoic of Utah and adjacent Idaho were undertaken in the late 1980s in conjunction with Ph.D. student Marjorie Levy (now at Chevron Petroleum Technology Company). Co-editor Paul K. Link (Idaho State University) has collaborated on Proterozoic geology in the western U.S. over a span of 20 years. Todd A. Ehlers, Marjorie A. Chan, Linda E. Sohl and Paul Link are thanked for helpful reviews of the manuscript. This guide was stimulated by the activities of IGCP Project 320 (1991–1995) and the International Commission on Stratigraphy Working Group on the Terminal Proterozoic System.

REFERENCES CITED

Aitken, J.D., 1991, The Ice Brook Formation and post-Rapitan, Late Proterozoic glaciation, Mackenzie Mountains, Northwest Territories: Geological Survey of Canada Bulletin 404, 43 p.

Allmendinger, R.W., 1992, Fold and thrust tectonics of the western United States exclusive of the accreted terranes, *in* Burchfiel, B.C., Lipman, P.W., and Zoback, M.L., eds., The Cordilleran Orogen: Conterminous U.S.: Boulder, Colorado, Geological Society of America, The Geology of North America, v. G-3, p. 583–607.

Armin, R.A., and Mayer, L., 1983, Subsidence analysis of the Cordilleran miogeocline: Implications for timing of late Proterozoic rifting and amount of extension: Geology, v. 11, p. 702–705.

Armstrong, R.L., 1968, Sevier orogenic belt in Nevada and Utah: Geological Society of America Bulletin, v. 79, p. 429–458.

Blick, N., 1979, Stratigraphic, structural and paleogeographic interpretation of upper Proterozoic glaciogenic rocks in the Sevier orogenic belt, northwestern Utah: Santa Barbara, University of California, Ph.D. dissertation, 633 p.

Bond, G.C., and Kominz, M.A., 1984, Construction of tectonic subsidence curves for the early Paleozoic miogeocline, southern Canadian Rocky Mountains: Implications for subsidence mechanisms, age of breakup, and crustal thinning: Geological Society of America Bulletin, v. 95, p. 155–173.

Bond, G.C., Christie-Blick, N., Kominz, M.A., and Devlin, W.J., 1985, An early Cambrian rift to post-rift transition in the Cordillera of western North America: Nature, v. 316, p. 742–745.

Bond, G.C., Kominz, M.A., and Devlin, W.J., 1983, Thermal subsidence and eustasy in the Lower Paleozoic miogeocline of western North America: Nature, v. 306, p. 775–779.

Bruhn, R.L., Picard, M.D., and Isby, J.S., 1986, Tectonics and sedimentology of Uinta Arch, western Uinta Mountains and Uinta Basin, *in* Peter- son, J.A., ed., Paleotectonics and sedimentation in the Rocky Moun- tain region, United States: American Association Petroleum Geologists Memoir 41, p. 333–352.

Bryant, B., 1990, Geologic map of the Salt Lake City 30′x60′ quadrangle, north-central Utah, and Uinta County, Wyoming: U.S. Geological Survey, Map I-1944, scale 1:100,000.

Burchfiel, B.C., and Davis, G.A., 1975, Nature and controls of Cordilleran orogenesis, western United States: extensions of an earlier synthesis: American Journal of Science, v. 275-A, p. 363–396.

Christie-Blick, N., 1982, Upper Proterozoic and Lower Cambrian rocks of the Sheeprock Mountains, Utah: regional correlation and significance: Geological Society of America Bulletin, v. 93, p. 735–750.

Christie-Blick, N., 1983a, Structural geology of the southern Sheeprock Mountains, Utah: regional significance, *in* Miller, D.M., Todd, V.R., and Howard, K.A., eds., Tectonic and stratigraphic studies in the eastern Great Basin: Geological Society of America Memoir 157, p. 101–124.

Christie-Blick, N., 1983b, Glacial-marine and subglacial sedimentation, Upper Proterozoic Mineral Fork Formation, Utah, *in* Molnia, B.F., ed., Glacial-marine sedimentation: New York, Plenum Press, p. 703–776.

Christie-Blick, N., 1991, Onlap, offlap, and the origin of unconformity-bounded depositional sequences: Marine Geology, v. 97, p. 35–56.

Christie-Blick, N., and Driscoll, N.W., 1995, Sequence stratigraphy: Annual Review of Earth and Planetary Sciences, v. 23, p. 451–478.

Christie-Blick, N., and Levy, M., 1985, A new approach to time correlation in Proterozoic rocks: sequence boundaries in the Brigham Group, Utah: Geological Society of America Abstracts with Programs, v. 17, p. 546.

Christie-Blick, N., and Levy, M., 1989a, Stratigraphic and tectonic framework of Upper Proterozoic and Cambrian rocks in the western United States, *in* Christie-Blick, N., and Levy, M., eds., Late Proterozoic and Cambrian Tectonics, Sedimentation, and Record of Metazoan Radiation in the Western United States: Washington, D.C., American Geophysical Union, 28th International Geological Congress, Field Trip Guidebook T331, p. 7–21.

Christie-Blick, N., and Levy, M., 1989b, Concepts of sequence stratigraphy, with examples from strata of Late Proterozoic and Cambrian age in the western United States, *in* Christie-Blick, N., and Levy, M., eds., Late Proterozoic and Cambrian Tectonics, Sedimentation, and Record of Metazoan Radiation in the Western United States: Washington, D.C., American Geophysical Union, 28th International Geological Congress, Field Trip Guidebook T331, p. 23–37.

Christie-Blick, N., and Link, P.K., 1988, Glacial-marine sedimentation, Mineral Fork Formation (Late Proterozoic), Utah, *in* Holden, G.S., ed., Geological Society of America field trip guidebook, centennial meeting—Denver, Colorado: Colorado School of Mines Professional Contributions, No. 12, p. 259–274.

Christie-Blick, N., Dyson, I.A., and von der Borch, C.C., 1995, Sequence stratigraphy and the interpretation of Neoproterozoic earth history. Precambrian Research, v. 73, p. 3–26.

Christie-Blick, N., Grotzinger, J.P., and von der Borch, C C., 1988, Sequence stratigraphy in Proterozoic successions. Geology, v. 16, p 100–104.

Christie-Blick, N , Mount, J.F., Levy, M., Signor, P.W., and Link, P.K., 1989, Description of stops, *in* Christie-Blick, N., and Levy, M., eds., Late Proterozoic and Cambrian Tectonics, Sedimentation, and Record of Metazoan Radiation in the Western United States. Washington, D.C., American Geophysical Union, 28th International Geological Congress, Field Trip Guidebook T331, p. 55–99.

Cohenour, R.E , 1959, Sheeprock Mountains, Tooele and Juab Counties Precambrian and Paleozoic stratigraphy, igneous rocks, structure, geomorphology, and economic geology. Utah Geological and Mineralogical Survey Bulletin 63, 201 p.

Crabtree, R D., 1981, Subglacial morphology in northern Palmer Land, Antarctic Peninsula: Annals of Glaciology, v. 2, p. 17–22.

Crittenden, M.D., Jr., 1965a, Geology of the Mount Aire quadrangle, Salt Lake County, Utah: U.S. Geological Survey, Map GQ-379, scale 1·24,000

Crittenden, M.D , Jr , 1965b, Geology of the Draper quadrangle, Utah· U.S. Geological Survey, Map GQ-377, scale 1·24,000

Crittenden, M.D., Jr., 1965c, Geology of the Sugar House quadrangle, Salt Lake County, Utah. U.S. Geological Survey, Map GQ-380, scale 1.24,000.

Crittenden, M.D., Jr., 1965d, Geology of the Dromedary Peak quadrangle, Utah. U.S. Geological Survey, Map GQ-378, scale 1.24,000.

Crittenden, M.D., Jr., 1976, Stratigraphic and structural setting of the Cottonwood area, Utah, *in* Hill, J.G , ed , Symposium on geology of the Cordilleran hingeline: Denver, Colorado, Rocky Mountain Association of Geologists, p. 363–379.

Crittenden, M.D., Jr, Christie-Blick, N., and Link, P.K., 1983, Evidence for two pulses of glaciation during the Late Proterozoic in northern Utah and southeastern Idaho: Geological Society of America Bulletin, v. 94, p. 437–450.

Crittenden, M D , Jr., Schaeffer, F.E , Trimble, D.E., and Woodward, L.A , 1971, Nomenclature and correlation of some upper Precambrian and basal Cambrian sequences in western Utah and southeastern Idaho. Geological Society of America Bulletin, v 82, p. 581–602

Crittenden, M.D., Sharp, B.J., and Calkins, F.C., 1952, Geology of the Wasatch Mountains east of Salt Lake City, Parleys Canyon to Traverse Range, *in* Marsell, R.E., ed., Geology of the central Wasatch Mountains: Utah Geological and Mineralogical Survey Guidebook to the Geology of Utah, No. 8, p 1–37

Crittenden, M.D., Jr., and Wallace, C.A , 1973, Possible equivalents of the Belt Supergroup in Utah, *in* Belt Symposium, v. 1. Moscow, University of Idaho, Idaho Bureau of Mines and Geology, p 116–138

Denton, G.H., Armstrong, R L., and Stuiver, M , 1971, The late Cenozoic glacial history of Antarctica, *in* Turekian, K.K., ed., Late Cenozoic Glacial Ages: New Haven, Connecticut, Yale University Press, p 267–306.

Diehl, P., 1976, Stratigraphy and sedimentology of the Wood Canyon Formation, Death Valley area, California, *in* Troxel, B.W., and Wright, L.A., eds , Geologic features of Death Valley, California. California Division of Mines and Geology Special Report 106, p. 51–62.

Eaton, G.P., 1982, The Basin and Range province—origin and tectonic significance. Annual Review of Earth and Planetary Sciences, v. 10, p. 409–440

Ehlers, T.A., Chan, M.A., and Link, P.K., 1997, Proterozoic tidal, glacial, and fluvial sedimentation in Big Cottonwood Canyon, Utah: Brigham Young University Geology Studies, v. 23 (this volume).

Eisbacher, G.H., 1985, Late Proterozoic rifting, glacial sedimentation, and sedimentary cycles in the light of Windermere deposition, western Canada Palaeogeography, Palaeoclimatology, Palaeoecology, v 51, p. 231–254

Granger, A.E., and Sharp, B.J., 1952, Geology of the Wasatch Mountains east of Salt Lake City (City Creek to Parleys Canyon), *in* Marsell, R E , ed., Geology of the central Wasatch Mountains: Utah Geological and Mineralogical Survey Guidebook to the Geology of Utah, No. 8, p. 1–37.

Groff, S.L., 1959, Geology of the West Tintic Range and vicinity, Tooele and Juab Counties, Utah: Salt Lake City, University of Utah, Ph D thesis, 183 p.

Grotzinger, J.P., Bowring, S.A., Saylor, B.Z., and Kaufman, A J , 1995, Biostratigraphic and geochronologic constraints on early animal evolution. Science, v. 270, p. 598–604.

Harper, G.D., and Link, P.K., 1986, Geochemistry of Upper Proterozoic rift-related volcanics, northern Utah and southeastern Idaho. Geology, v 14, p. 864–867.

Hein, F.J., and McMechan, M.E., 1994, Proterozoic and Lower Cambrian strata of the western Canada sedimentary basin, *in* Mossop, G D., and Shetsen, I., eds., Geological Atlas of the Western Canada Sedimentary Basin Calgary, Alberta, Canadian Society of Petroleum Geologists, p. 57–67.

Hintze, L.F., 1988, Geologic History of Utah. Brigham Young University Geology Studies, Special Publication 7, 202 p.

Hoffman, P.F, 1991, Did the breakout of Laurentia turn Gondwanaland inside-out?: Science, v. 252, p. 1409–1412.

James, L.P., 1979, Geology, ore deposits, and history of the Big Cottonwood Mining District, Salt Lake Couty, Utah. Utah Geological and Mineral Survey, Bulletin 114, 98 p

Levy, M., and Christie-Blick, N., 1989, Pre-Mesozoic palinspastic reconstruction of the eastern Great Basin (western United States): Science, v. 245, p. 1454–1462.

Levy, M., and Christie-Blick, N., 1991a, Late Proterozoic paleogeography of the eastern Great Basin, *in* Cooper, J.D., and Stevens, C.H , eds., Paleozoic Paleogeography of the Western United States—II. Los Angeles, California, Pacific Section, Society of Economic Paleontologists and Mineralogists, v. 1, p. 371–386.

Levy, M., and Christie-Blick, N., 1991b, Tectonic subsidence of the early Paleozoic passive continental margin in eastern California and southern Nevada: Geological Society of America Bulletin, v. 103, p. 1590–1606.

Levy, M., Christie-Blick, N., and Link, P.K., 1994, Neoproterozoic incised valleys of the eastern Great Basin, Utah and Idaho. Fluvial response to changes in depositional base level, *in* Dalrymple, R.W, Boyd, R , Zaitlin, B A., eds., Incised-Valley Systems. Origin and Sedimentary Sequences. SEPM (Society for Sedimentary Geology) Special Publication No. 51, p. 369–382.

Levy, M E., 1991, Late Proterozoic and Early Cambrian sedimentation, sequence stratigraphy, and tectonic evolution of the eastern Great Basin: New York, Columbia University, Ph D dissertation, 380 p

Link, P.K., 1984, Comment on "Subsidence analysis of the Cordilleran miogeocline: Implications for timing of Late Proterozoic rifting and amount of extension": Geology, v 12, p 699.

Link, P.K., Christie-Blick, N., Devlin, W.J., Elston, D.P., Horodyski, R.J , Levy, M., Miller, J.M.G., Pearson, R.C., Prave, A., Stewart, J.H , Winston, D., Wright, L.A., and Wrucke, C.T., 1993, Middle and Late Proterozoic stratified rocks of the western U.S. Cordillera, Colorado Plateau, and Basin and Range province, *in* Reed, J.C , Jr, Bickford, M.E., Houston, R.S., Link, P.K., Rankin, D.W, Sims, P.K , and Van Schmus, W.R., eds., Precambrian: Conterminous U S Boulder, Colorado, Geological Society of America, The Geology of North America, v. C-2, p. 463–595

Link, P.K., Jansen, S.T., Halimdihardja, P., Lande, A., and Zahn, P., 1987, Stratigraphy of the Brigham Group (Late Proterozoic-Cambrian), Bannock, Portneuf, and Bear River Ranges, southeastern Idaho, *in* Miller, W.R., ed., The thrust belt revisited: Wyoming Geological Association, 38th annual field conference guidebook, p. 133–148.

Link, P.K., Miller, J.M.G., and Christie-Blick, N., 1994, Glacial-marine facies in a continental rift environment. Neoproterozoic rocks of the western United States Cordillera, *in* Deynoux, M., Miller, J.M.G., Domack, E.W., Eyles, N, Fairchild, I J., and Young, G.M., eds., Earth's Glacial Record Cambridge, Cambridge University Press, p 29–46

Lochman-Balk, C., 1976, The Cambrian section of the central Wasatch Mountains, *in* Hill, J.G., ed., Symposium on geology of the Cordilleran hingeline· Denver, Colorado, Rocky Mountain Association of Geologists, p. 103–108.

Loucks, R.G., and Sarg, J F, eds., 1993, Carbonate Sequence Stratigraphy· Recent Developments and Applications. American Association of Petroleum Geologists Memoir 57, 545 p.

Miller, K G , Mountain, G.S , the Leg 150 Shipboard Party, and Members of the New Jersey Coastal Plain Drilling Project, 1996, Drilling and dating New Jersey Oligocene-Miocene sequences. Ice volume, global sea level, and Exxon records: Science, v 271, p. 1092–1095.

Mitchum R M., Jr., 1977, Seismic stratigraphy and global changes of sea level, part 11: Glossary of terms used in seismic stratigraphy, *in* Payton, C E., ed., Seismic Stratigraphy—Applications to Hydrocarbon Exploration: American Association of Petroleum Geologists Memoir 26, p. 205–212.

Morris, H.T., 1977, Geologic map and sections of the Furner Ridge quadrangle, Juab County, Utah: U.S. Geological Survey, Map I-1045, scale 1.24,000

Morris, H.T., 1983, Interrelations of thrust and transcurrent faults in the central Sevier orogenic belt near Leamington, Utah, *in* Miller, D.M., Todd, V.R., and Howard, K.A., eds., Tectonic and stratigraphic studies in the eastern Great Basin: Geological Society of America Memoir 157, p. 75–81

Morris, H.T, and Kopf, R.W, 1967, Breccia pipes in the West Tintic and Sheeprock Mountains, Utah: U S. Geological Survey Professional Paper 575-C, p. 66–71

Morris, H.T., and Kopf, R.W., 1970a, Preliminary geologic map and cross section of the Cherry Creek quadrangle and adjacent part of the Dutch Peak quadrangle, Juab County, Utah. U.S. Geological Survey Open-File Map, scale 1:24,000.

Morris, H.T, and Kopf, R.W, 1970b, Preliminary geologic map and cross section of the Maple Peak quadrangle and adjacent part of the Sabie Mountain quadrangle, Juab County, Utah U.S. Geological Survey Open-File Map, scale 1:24,000.

Mount, J.F., 1982, Storm-surge-ebb origin of hummocky cross-stratified units of the Andrews Mountain Member, Campito Formation (Lower Cambrian), White-Inyo Mountains, eastern California. Journal of Sedimentary Petrology, v. 52, p. 941–958.

Mukul, M , and Mitra, G , 1994, Deformation history of the Sheeprock thrust sheet, Utah New interpretations based on structural analysis Geological Society of America Abstracts with Programs, v. 26, No 6, p. 55.

Oberbeck, V.R., Hörz, F., and Bunch, T., 1994, Impacts, tillites, and the breakup of Gondwanaland· A second reply· Journal of Geology, v. 102, p. 485–489.

Ojakangas, R.W., and Matsch, C.L., 1980, Upper Precambrian (Eocambrian) Mineral Fork Tillite of Utah. A continental glacial and glaciomarine sequence. Geological Society of America Bulletin, Part I, v. 91, p. 495–501.

Posamentier, H W., and James, D P., 1993, An overview of sequence-stratigraphic concepts. uses and abuses, *in* Posamentier, H.W., Summerhayes, C.P., Haq, B.U., and Allen, G.P., eds., Sequence Stratigraphy and Facies Associations. International Association of Sedimentologists Special Publication 18, p 3–18.

Posamentier, H.W., Jervey, M.T., and Vail, P.R., 1988, Eustatic controls on clastic deposition I—conceptual framework, *in* Wilgus, C.K., Hastings, B.S., Kendall, C.G.St.C., Posamentier, H.W., Ross, C.A., and van Wagoner, J.C., eds., Sea-level changes. an integrated approach: Society of Economic Paleontologists and Mineralogists Special Publication No. 42, p. 109–124.

Powell, C.McA., Preiss, W.V., Gatehouse, C.G., Krapez, B., and Li, Z.X., 1994, South Australian record of a Rodinian eopicontinental basin and its mid-Neoproterozoic breakup (~700 Ma) to form the Palaeo-Pacific Ocean Tectonophysics, v. 237, p. 113–140.

Rainbird, R.H., Jefferson, C.W., and Young, G.M., 1996, The early Neoproterozoic sedimentary Succession B of northwestern Laurentia: Correlations and paleogeographic significance: Geological Society of America Bulletin, v. 108, p 454–470.

Ross, G.M., 1991, Tectonic setting of the Windermere Supergroup revisited: Geology, v. 19, p. 1125–1128.

Ross, G.M., Bloch, J.D., and Krouse, H.R., 1995, Neoproterozoic strata of the southern Canadian Cordillera and the isotopic evolution of seawater sulfate. Precambrian Research, v 73, p. 71–99.

Schmidt, P.W., and Williams, G.E., 1995, The Neoproterozoic climatic paradox: Equatorial palaeolatitude for Marinoan glaciation near sea level in South Australia. Earth and Planetary Science Letters, v. 134, p. 107–124.

Sears, J W., Graff, P.J., and Holden, G S., 1982, Tectonic evolution of lower Proterozoic rocks, Uinta Mountains, Utah and Colorado: Geological Society of America Bulletin, v. 93, p. 990–997.

Smith, R.B., and Bruhn, R.L., 1984, Intraplate extensional tectonics of the eastern Basin-Range: inferences on structural style from seismic reflection data, regional tectonics, and thermal-mechanical models of brittle-ductile deformation: Journal of Geophysical Research, v. 89, p. 5733–5762

Stewart, J.H., 1970, Upper Precambrian and Lower Cambrian strata in the southern Great Basin, California and Nevada· U.S. Geological Survey Professional Paper 620, 206 p.

Stewart, J H., 1972, Initial deposits in the Cordilleran geosyncline. evidence of a Late Precambrian (<850 m.y.) continental separation: Geological Society of America Bulletin, v. 83, p. 1345–1360.

Stewart, J.H , 1976, Late Precambrian evolution of North America: plate tectonics implication: Geology, v. 4, p. 11–15.

Stewart, J.H., 1978, Basin and Range structure in western North America—a review, *in* Smith, R.B., and Eaton, G P, eds., Cenozoic tectonics and regional geophysics of the western Cordillera. Geological Society of America Memoir 152, p. 1–31.

Stewart, J.H., 1982, Regional relations of Proterozoic Z and Lower Cambrian rocks in the western United States and northern Mexico, *in* Cooper, J.D., Troxel, B.W., and Wright, L.A., eds., Geology of selected areas in the San Bernardino Mountains, western Mojave Desert, and southern Great Basin, California: Geological Society of America Cordilleran section meeting, Anaheim, California, volume and guidebook, p. 171–186.

Stewart, J.H., and Poole, F.G., 1974, Lower Paleozoic and uppermost Precambrian Cordilleran miogeocline, Great Basin, western United States, *in* Dickinson, W.R., ed., Tectonics and sedimentation: Society of Economic Paleontologists and Mineralogists Special Publication No. 22, p 28–57.

Stewart, J.H , and Suczek, C A., 1977, Cambrian and latest Precambrian paleogeography and tectonics in the western United States, *in* Stewart, J H , Stevens, C H., and Fritsche, A.E., eds., Paleozoic paleogeography of the western United States: Society of Economic Paleontologists and Mineralogists, Pacific section, Pacific Coast Paleogeography Symposium 1, p. 1–17.

Torsvik, T.H , Smethurst, M.A., Meert, J.G., Van der Voo, R., McKerrow, W.S., Brasier, M.D., Sturt, B.A., and Walderhaug, H.J., 1996, Continental break-up and collision in the Neoproterozoic and

Palaeozoic—A tale of Baltica and Laurentia: Earth-Science Reviews, v. 40, p. 229–258.

Vail, P.R., 1987, Seismic stratigraphy interpretation using sequence stratigraphy. Part I: Seismic stratigraphy intepretation procedure, *in* Bally, A.W, ed., Atlas of seismic stratigraphy· American Association of Petroleum Geologists Studies in Geology No. 27, v. 1, p. 1–10.

Van Wagoner, J.C., Mitchum, R.M , Campion, K M , and Rahmanian, V.D., 1990, Siliciclastic Sequence Stratigraphy in Well Logs, Cores, and Outcrops. Concepts for High-Resolution Correlation of Time and Facies. American Association of Petroleum Geologists Methods in Exploration Series, No. 7, 55 p.

Van Wagoner, J.C., Posamentier, H.W., Mitchum, R.M., Vail, P.R., Sarg, J.F., Loutit, T.S., and Hardenbol, J., 1988, An overview of the fundamentals of sequence stratigraphy and key definitions, *in* Wilgus, C.K., Hastings, B.S., Kendall, C.G.St.C , Posamentier, H.W., Ross, C.A , and van Wagoner, J.C., eds., Sea-level changes: an integrated approach· Society of Economic Paleontologists and Mineralogists Special Publication No. 42, p. 39–45.

Veevers, J.J , Walter, M.R., and Scheibner, E., 1997, Neoproterozoic tectonics of Australia-Antarctica and Laurentia and the 560 Ma birth of the Pacific Ocean reflect the 400 m y. Pangean Supercycle. Journal of Geology, v. 105, p. 225–242

Weimer, P., and Posamentier, H W., eds., 1993, Siliciclastic Sequence Stratigraphy. Recent Developments and Applications American Association of Petroleum Geologists Memoir 58, 492 p.

Wernicke, B., 1992, Cenozoic extensional tectonics of the U.S. Cordillera, *in* Burchfiel, B.C., Lipman, P.W., and Zoback, M.L , eds., The Cordilleran Orogen. Conterminous U.S.. Boulder, Colorado, Geological Society of America, The Geology of North America, v. G-3, p. 553–581.

Young, G.M., 1982, The late Proterozoic Tindir Group, east-central Alaska Evolution of a continental margin. Geological Society of America Bulletin, v. 93, p. 759–783.