# Chapter 38

# Neoproterozoic strata of southeastern Idaho and Utah: record of Cryogenian rifting and glaciation

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**Abstract:** Neoproterozoic strata in southeastern Idaho and Utah include the <766 Ma Uinta Mountain Group and Big Cottonwood Formation (Fm.) deposited in an east-trending rift basin and, to the west, the lower part of a westward-thickening rift to passive-margin succession that initiated *c.* 720 Ma. The latter contains a lower diamictite and volcanic succession, with a complex stratigraphic interval of Cryogenian marine glacial deposits (Pocatello and Mineral Fork formations and correlatives). This is overlain by a mostly terrigenous succession of <667 Ma strata assigned to the upper member of the Pocatello Fm. and Brigham Group in southeastern Idaho, to the Kelley Canyon Fm. and Brigham Group in northern and western Utah, and to the McCoy Creek Group and Prospect Mountain Quartzite in adjacent Nevada. Although the Brigham Group and correlative deposits contain no direct evidence for glaciation, widely developed, though stratigraphically restricted, incised valleys, with erosional relief from a few metres to as much as 160 m, are inferred to represent subsequent times of Cryogenian glacially lowered sea level. Overall interpretations of the stratigraphy and sedimentology of these rocks have changed little in the past 10–15 years. The most important recent advances relate to U–Pb geochronology. In strata that lie unconformably below demonstrable glacial deposits, the lower Uinta Mountain Group (formerly thought to be *c.* 900 Ma) contains populations of detrital zircons as young as 766 ± 5 Ma. Cryogenian magmatism north of the Snake River Plain in central Idaho is recognized near House Mountain, east of Boise at *c.* 725 ± 5 Ma, in the Pioneer Mountains Core Complex at about 695 Ma, and in central and east-central Idaho at 685–650 Ma. Clasts interpreted to be from the rift-related Bannock Volcanic Member of the Pocatello Fm. are dated at 717 ± 4 Ma and 701 ± 4 Ma. The overlying diamictite-bearing Scout Mountain Member contains a mafic lapilli tuff near the base (686 ± 4 Ma) and a reworked fallout tuff near the top (667 ± 5 Ma). Strongly negative C-isotope data have been obtained from some of the carbonate rocks, although the latter constitute only a small fraction of the succession. Palaeomagnetic data are available only for the Uinta Mountain Group, and suggest an equatorial palaeolatitude.

Neoproterozoic glaciogenic rocks, locally in excess of 1 km thick, are exposed widely but discontinuously in southeastern Idaho, and northern and western Utah (Fig. 38.1). The deposits are assigned to a plethora of local formal and informal stratigraphic units, for historical reasons and because of their varied expression. Names used for commonly correlated diamictite-bearing units include Pocatello Fm. in Idaho, and Mineral Fork Fm., Sheeprock Group, formation of Perry Canyon, Trout Creek sequence (units 3 and 5), and Horse Canyon Fm. in Utah (Fig. 38.2). Crittenden *et al.* (1971, 1983) and Link *et al.* (1993, 1994) provide regional reviews of diamictites and their interpretation. Important papers for specific locations include Blackwelder (1932), Crittenden *et al.* (1952), Ojakangas & Matsch (1980), Blick (1981), Christie-Blick (1982, 1983*a*, 1983*b*, 1985, 1997), Christie-Blick & Link (1988), Christie-Blick & Levy (1989) and Yonkee *et al.* (2000*a*) for northern and west-central Utah; Ludlum (1942), Trimble (1976), Link (1981, 1983, 1987; Link *et al.* 2005) for the Pocatello area of Idaho; and Misch & Hazzard (1962), Bick (1966) and Rodgers (1994) for westernmost Utah and eastern Nevada. Type localities are described by Misch & Hazzard (1962; Trout Creek sequence and McCoy Creek Group), Bick (1966; Horse Canyon Fm.), Crittenden *et al.* (1971; Kelley Canyon Fm., Brigham Group and Pocatello Fm.), Christie-Blick (1982, 1983*a*; Mineral Fork Fm. and Sheeprock Group), Crittenden *et al.* (1983; Pocatello Fm.), Link (1983, 1987; Pocatello Fm.) and Link *et al.* (1985; Brigham Group).

The stratigraphy and sedimentology of these strata were worked out primarily in the 1970s and early 1980s. The correlations first synthesized by Crittenden *et al.* (1971) were based on regional field mapping, and have largely stood the test of time as summarized in Link *et al.* (1993). More recent research has focused on U–Pb geochronology (Fanning & Link 2004, 2008) and geochemistry (Smith *et al.* 1994; Young 2002; Lorentz *et al.* 2004; Corsetti *et al.* 2007; Dehler *et al.* 2007).

## Structural framework

The rocks are thought to have accumulated in rift-related basins associated with the development of a passive continental margin in western North America between *c.* 665 Ma and *c.* 520 Ma, and in part atop erosional topography with as much as 900 m of local relief (Stewart 1972; Stewart & Suczek 1977; Bond *et al.* 1983, 1985; Christie-Blick & Levy 1989; Levy & Christie-Blick 1991*a*; Ross 1991; Christie-Blick 1983*a*, 1997). They crop out today within and along the eastern flank of the late Jurassic to early Cenozoic Cordilleran thrust-and-fold belt, and across the eastern edge of the late Cenozoic Basin and Range extensional province (Fig. 38.1; Armstrong & Oriel 1965; Armstrong 1968; Levy & Christie-Blick 1989; Allmendinger 1992; Wernicke 1992; DeCelles 2004; DeCelles & Coogan 2006). The structurally lowest thrust sheets encompassing thick glacial and associated deposits at the present level of exposure (generally on the eastern side of the thrust belt) belong to the Willard–Paris–Putnam system in northern Utah and southeastern Idaho, and to the Tintic–Sheeprock–Canyon Range system of west-central Utah (Levy & Christie-Blick 1989; DeCelles 2004; Fig. 38.1). For example, the Pocatello Fm. in the Bannock Range of southeastern Idaho and the formation of Perry Canyon near Ogden in the northern Wasatch Range of Utah represent a 200 km strike-parallel outcrop band within the Putnam-Paris thrust sheet. Although the rocks are generally foliated, with pervasive development of chlorite and locally biotite in the greenschist facies, detrital zircon geochronology and C-isotope studies have been successfully

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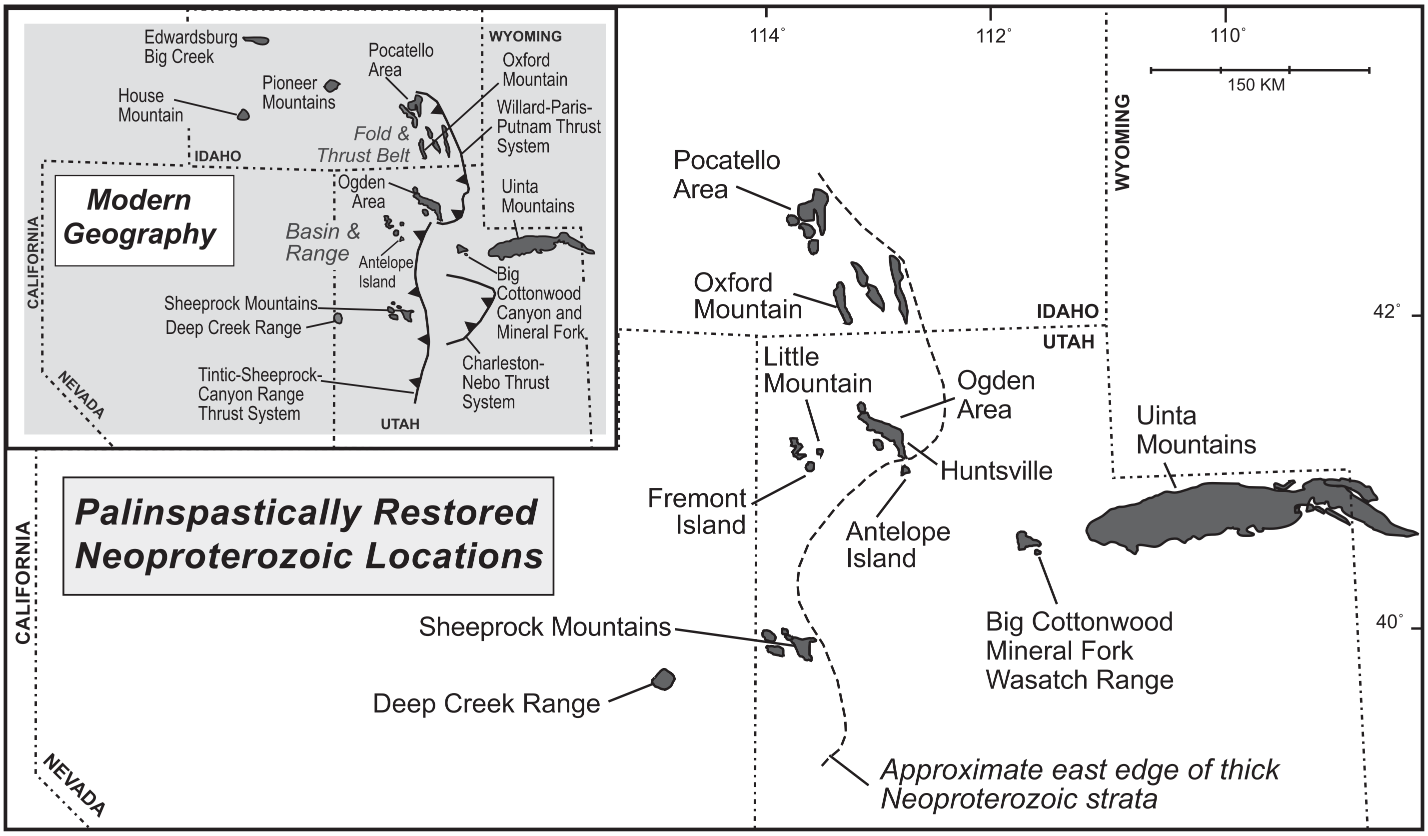


**Fig. 38.1.** Map showing areas of outcrop of Neoproterozoic rocks of Idaho and Utah, with key locations shown (after Link *et al.* 1994). Inset map is present geography. Main map is palinspastic reconstruction after Levy & Christie-Blick (1989).

conducted (Fanning & Link 2004, 2008; Lorentz *et al.* 2004; Corsetti *et al.* 2007). One or more generations of Neogene extensional faults cut all of these ranges. In many cases, therefore, rocks transported eastward during Cretaceous thrusting have been translated westward during Neogene extension (Levy & Christie-Blick 1989).

Detailed field studies suggest that primary sedimentary features and correlations are discernable through the deformation. In the structurally highest hinterland thrust sheet of the Deep Creek Range, close to the Utah–Nevada state line, garnet and staurolite grade rocks are present in the Trout Creek sequence (Nelson 1966; Rodgers 1994). East of there, but in the same thrust

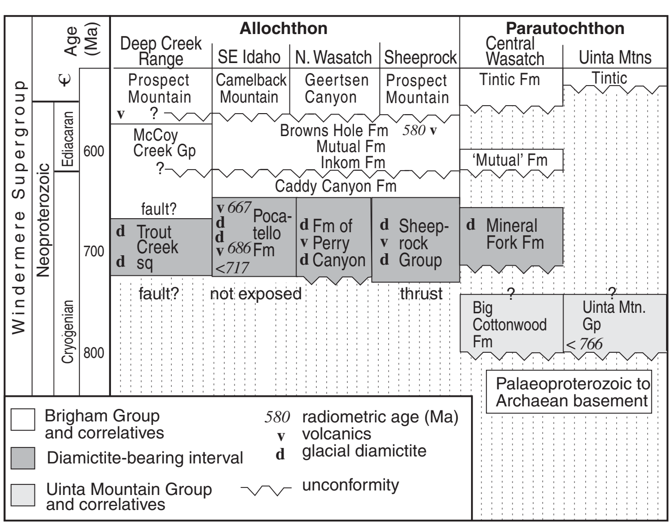


**Fig. 38.2.** Utah and Idaho Neoproterozoic correlation chart showing stratigraphic names, ages, and locations of diamictites, volcanic rocks and carbonate strata.

complex, the entire Neoproterozoic succession is overturned beneath the Pole Canyon thrust in the southern Sheeprock Mountains, over a lateral distance of *c.* 10 km (Christie-Blick 1983*b*). In the northern Wasatch Range, the formation of Perry Canyon crops out in an east-vergent overturned fold above the Willard thrust fault. At Portneuf Narrows, SE of Pocatello, the type section of the Pocatello Fm. is exposed in an overturned fold, cut by a Cretaceous tear fault.

Neoproterozoic rocks of the central Wasatch Range and Uinta Mountains are parautochthonous with respect to cratonic North America, having been displaced eastward no more than a few kilometres by mostly blind structures (Bruhn *et al.* 1986). In the Uinta Mountain Group some strata are basically unmetamorphosed and retain organic carbon (Dehler & Sprinkel 2005). The Big Cottonwood Fm. in the Wasatch Range is at greenschist facies. The least deformed and metamorphosed glacial deposits are found in the Mineral Fork Fm. of the central Wasatch parautochthon, except within the aureole adjacent to Oligocene stocks (Christie-Blick 1983*a*).

## Stratigraphy

Neoproterozoic and lower Cambrian, predominantly siliciclastic rocks in southeastern Idaho and adjacent Utah are divisible into three intervals. Pre-glacial deposits are best represented by locally conglomeratic sandstone and siltstone of the Uinta Mountain Group in the Uinta Mountains (Figs 38.2 & 38.3) and by comparable quartzite and argillite of the Big Cottonwood Fm. in the central Wasatch Range. Glacial and associated deposits are represented by the Pocatello Fm. and correlatives (Figs 38.2 & 38.4; Crittenden *et al.* 1971, 1983). Terminal Neoproterozoic to lower Cambrian quartzite, minor siltstone and minor carbonate are 'post-glacial' in terms of preserved facies. However, they are relevant to the theme of this volume because they contain an indirect record of sea-level change (incised valleys) that may reflect the

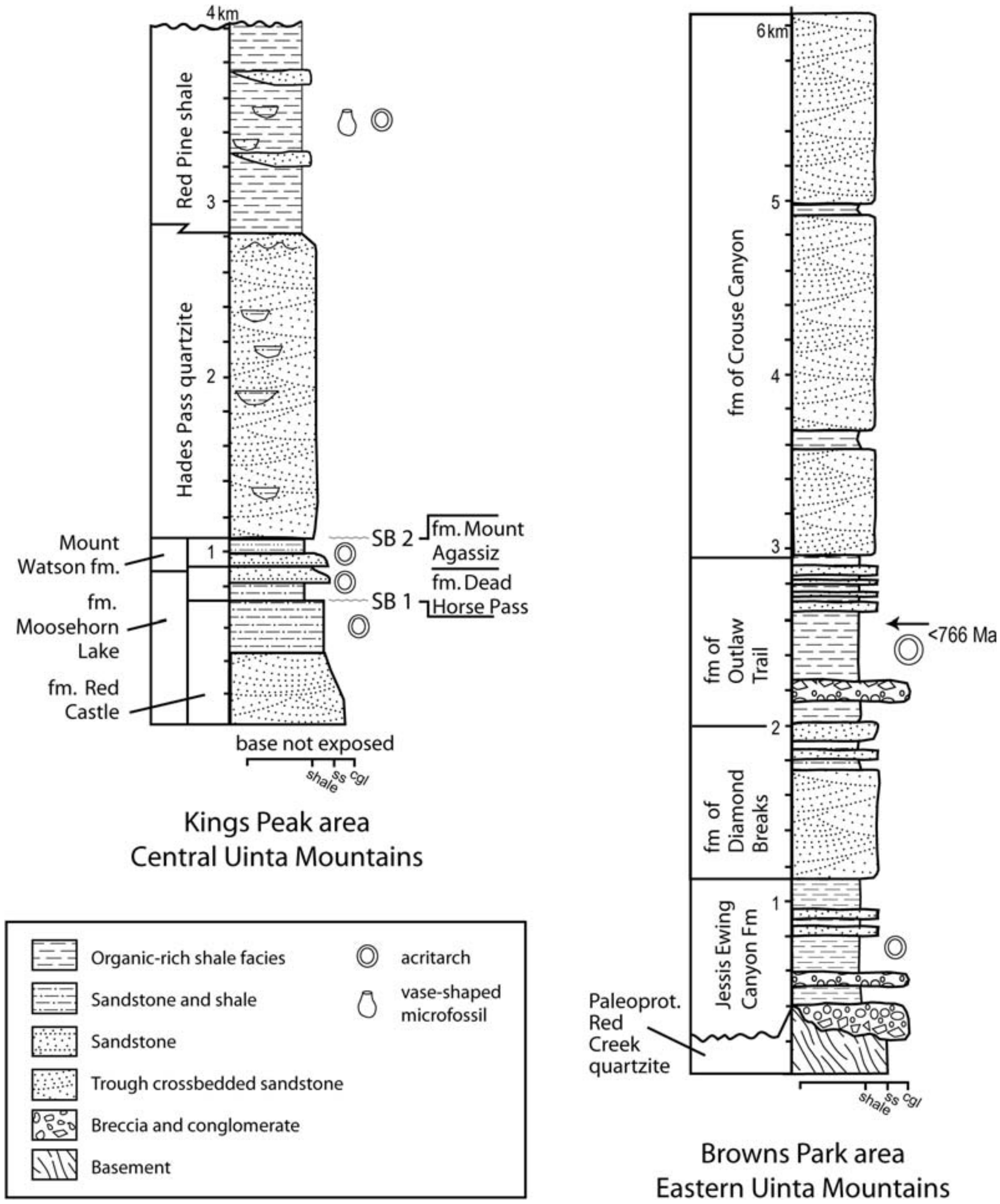


**Fig. 38.3.** Stratigraphy of the Uinta Mountain Group in the central Uinta Mountains (Kings Peak quadrangle) and in the eastern range near Brown's Park (after Dehler *et al.* 2007, and E. M. Kingsbury, Idaho State University, pers. comm. 2008). SB, sequence boundary.

waxing and waning of ice sheets elsewhere on the planet. The rocks are assigned in most places to the Brigham Group (Figs 38.2 & 38.5; Crittenden *et al.* 1971; Christie-Blick 1982; Link *et al.* 1985). Correlative strata in the Deep Creek Range are the McCoy Creek Group and overlying Prospect Mountain Quartzite (Fig. 38.2; Misch & Hazzard 1962). All of these Neoproterozoic and lower Cambrian strata are broadly equivalent to the Windermere Supergroup of Washington and western Canada and to the Kingston Peak Fm. and overlying rocks in Death Valley (Link *et al.* 1993; Lund *et al.* 2003, 2011; Hoffman & Halverson 2011; Mrofka & Kennedy 2011; Petterson *et al.* 2011; Smith *et al.* 2011).

### *Uinta Mountain Group and Big Cottonwood Fm.*

The Cryogenian but pre-glacial Uinta Mountain Group consists of 4–7 km of pervasively cross-bedded arkosic and quartzose sandstone, with subordinate medial and upper intervals of mudrock (Fig. 38.3; Hansen 1965; Sanderson & Wiley 1986; Stone 1993; Link *et al.* 1993; Dehler & Sprinkel 2005; Dehler *et al.* 2010). The Big Cottonwood Fm. is of similar thickness (5 km) and lithology (interstratified quartzite and argillite), but is more strongly folded and less easily studied than the Uinta Mountain Group in the steep-sided Wasatch Range canyons in which it is exposed. The most distinctive attribute of these apparently non-descript rocks is the presence of tidal rhythmites (Chan *et al.* 1994; Ehlers *et al.* 1997; Ehlers & Chan 1999). Direct correlation between the Uinta Mountain Group and Big Cottonwood Fm. is not possible because younger rocks intervene between available exposures. However, the successions are thought to be of broadly the same age on the basis of gross lithostratigraphy and provenance (Condie *et al.* 2001; Dehler *et al.* 2007), a stratigraphic position below glacial deposits in the case of the latter, and the east–west alignment of the sedimentary basins in which the successions are thought to have accumulated (Christie-Blick 1997). Further, detrital zircons in both units show the same populations, with the youngest population in the Uinta Mountain Group at <766 Ma (Dehler *et al.* 2007).

### *Pocatello Fm. and correlative units*

The Pocatello Fm. of southeastern Idaho consists of a lower interval of mafic volcanic flows, fragmental volcanic rocks and minor

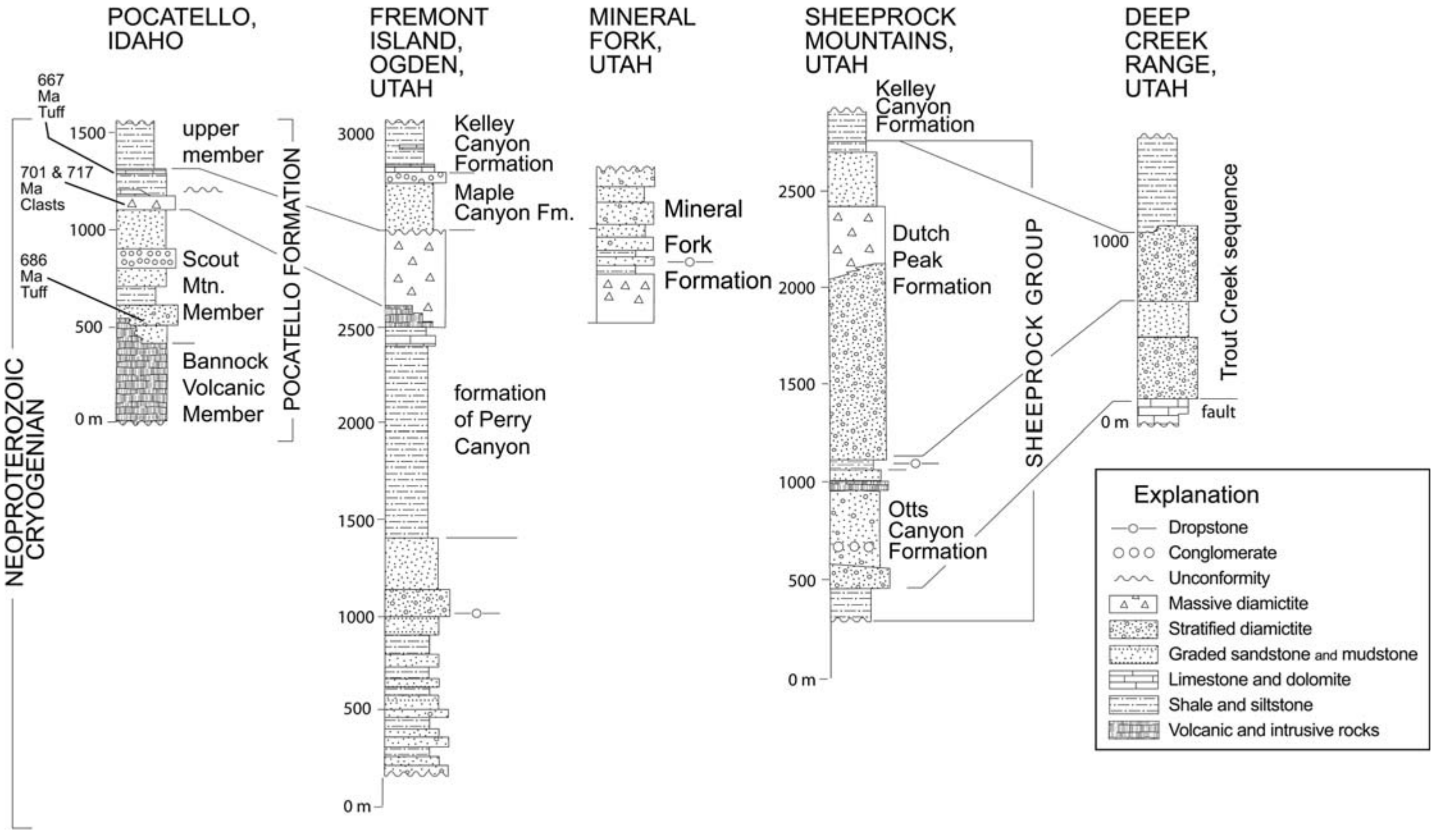


**Fig. 38.4.** Stratigraphy of diamictite and volcanic succession, southeastern Idaho and northern and western Utah (after Link *et al.* 1994). Available radiometric ages are from Fanning & Link (2004, 2008).

intrusives (Bannock Volcanic Member); a heterolithic medial unit (Scout Mountain Member) that includes two intervals of diamictite, cobble conglomerate, ferruginous sandstone and at least two zircon-bearing tuff beds; and an unnamed upper member consisting primarily of laminated shale or phyllite (Fig. 38.4; Ludlum 1942; Crittenden *et al.* 1971, 1983; Trimble 1976; Link 1981, 1983, 1987; Link *et al.* 2005; Rodgers *et al.* 2006). A thin laminated carbonate, locally resedimented as multiple layers of breccia, is present at the base of the upper member. The lower diamictite contains abundant intrabasinal volcanic clasts, whereas the upper contains clasts of quartzite (locally striated), granitic rocks and felsic volcanic rocks interpreted to have been derived from the subjacent Bannock Volcanic Member.

Lying unconformably on the Big Cottonwood Fm. in the vicinity of Mineral Fork in the central Wasatch Range, the Mineral Fork Fm. consists of as much as 800 m of diamictite, siltstone, sandstone and conglomerate (Fig. 38.4; Crittenden *et al.* 1952; Ojakangas & Matsch 1980; Christie-Blick 1983*a*, 1997; Link *et al.* 1994). The diamictite, which consists primarily of rounded clasts of quartzite, carbonate, other sedimentary rocks and minor igneous rocks in a sandy matrix, contains striated stones, lonestones in laminated host rock, and sheets and lenses of texturally mature sandstone. These strata are unconformably overlain in turn by quartzite of Mutual Fm. and Tintic Quartzite (Ediacaran to lower Cambrian). Comparable deposits, also assigned to the Mineral Fork Fm., are present in the Charleston-Nebo thrust sheet of the southern Wasatch Range (small outcrops not shown in Fig. 38.1), and in the parautochthon of Antelope Island. In the case of the latter, the Mineral Fork Fm. rests not on the Big Cottonwood Fm., but on gneisses of Archaean to Palaeoproterozoic age, and it is overlain more or less conformably by a well-developed laminated carbonate and argillite correlative with the upper member of the Pocatello Fm. at Pocatello (Kelley Canyon Fm.; Christie-Blick 1983*a*; Crittenden *et al.* 1983; Bryant 1988; Yonkee *et al.* 2000*a*). Diamictite at this location is coarse-grained, with clasts composed largely of gneiss.

The formation of Perry Canyon of northern Utah consists of more than 2 km of diamictite, greywacke, argillite, sandstone and mafic volcanic rocks (Crittenden *et al.* 1983). These strata crop out widely though discontinuously in the Willard thrust sheet (Ogden area, Huntsville, Little Mountain and Fremont Island in Fig. 38.1). Neither the base nor the top is exposed at most locations. North and east of Ogden, in the northern Wasatch Range, however, the Perry Canyon locally overlies the Palaeoproterozoic Facer Fm., and underlies a poorly developed laminated carbonate. The latter passes upward into several hundred metres of siltstone (Kelley Canyon Fm.; Crittenden *et al.* 1971). Diamictite clasts consist primarily of granitic rocks and quartzite (Crittenden *et al.* 1983).

More than 2 km of diamictite, greywacke, conglomerate, quartzite and argillite is present in the Otts Canyon Fm. and overlying Dutch Peak Fm. of the Sheeprock Group in the Sheeprock Mountains in west-central Utah (Christie-Blick 1982, 1997; Crittenden *et al.* 1983). The lower part of the Otts Canyon Fm., stratigraphically below all exposed diamictite, consists of phyllitic argillite. The uppermost part of the Otts Canyon Fm. is intruded by mafic sills. The Dutch Peak Fm. is overlain directly by shale of the Kelley Canyon Fm. without an intervening carbonate. Diamictite in the Otts Canyon Fm. consists predominantly of quartzite clasts in a phyllitic matrix (Christie-Blick 1982). Substantially thicker diamictite and greywacke of the Dutch Peak Fm. consist primarily of clasts of granitic rocks, carbonate and quartzite in a phyllitic to sandy matrix, with notable lateral variations in the relative abundance of clast types.

Diamictite is present in units 3 and 5 of the lower part of the Trout Creek sequence of Misch & Hazzard (1962; Horse Canyon Fm. of Bick 1966) in the Deep Creek Range on the Utah–Nevada state line (Rodgers 1994). Details of the stratigraphic succession are less well established than at other locations owing to structural complexity and higher metamorphic grade (Christie-Blick 1982).

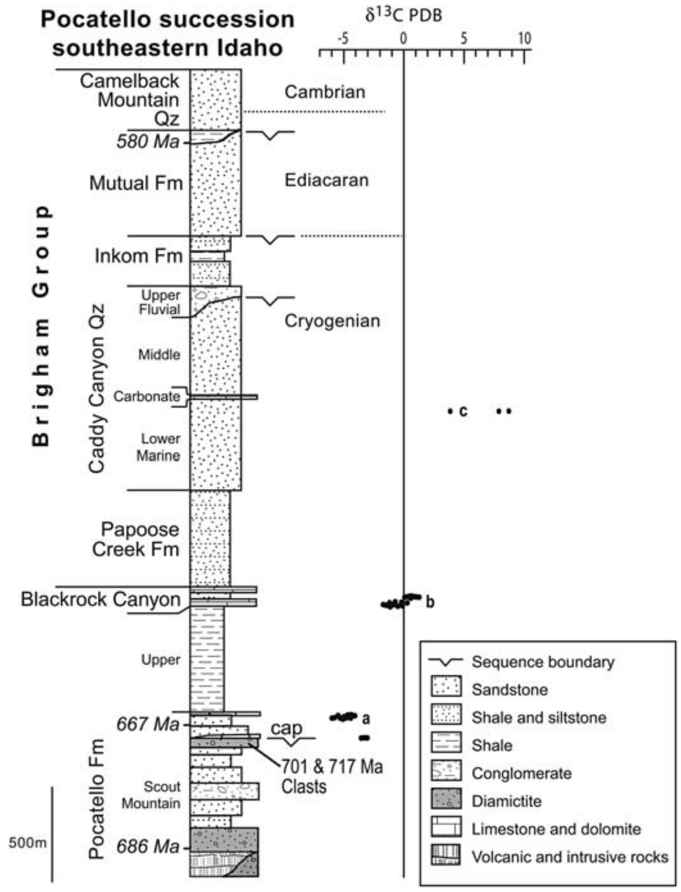


**Fig. 38.5.** Pocatello area (Pocatello Fm., Blackrock Canyon Limestone and Brigham Group) stratigraphy showing available C-isotope data (after Lorentz *et al.* 2004). In the C-isotope curve, interval 'a' is the cap carbonate that overlies the diamictite of the Scout Mountain Member, interval 'b' is the Blackrock Canyon Limestone, and interval 'c' is from thin dolomites within the marine Caddy Canyon Quartzite. Qz, quartzite.

### *Brigham Group*

In most areas of southeastern Idaho and northern Utah, many hundreds of metres of shale and argillite directly overlie glacial strata (upper member of the Pocatello Fm., Kelley Canyon Fm., and Trout Creek sequence unit 6). These pass upwards into a quartzose succession up to several kilometres thick of late Neoproterozoic (late Cryogenian and Ediacaran) to early Cambrian age. At most locations in southeastern Idaho and northern Utah, these strata are assigned to the Brigham Group (Crittenden *et al.* 1971; Christie-Blick 1982; Stewart 1982; Link *et al.* 1985, 1987, 1993; Levy & Christie-Blick 1991*b*). In westernmost Utah and eastern Nevada, the rocks are placed in the upper Trout Creek sequence, McCoy Creek Group and Prospect Mountain Quartzite (Misch & Hazzard 1962; Fig. 38.2).

The stratigraphy of the Brigham Group varies in detail, particularly in the distribution of carbonate rocks (e.g. Blackrock Canyon Limestone in Idaho) and volcanic rocks (e.g. Browns Hole Fm. in the northern Wasatch Range), and in the abundance of siltstone at the level of the Caddy Canyon Quartzite (Christie-Blick 1982). The interval that encompasses the upper part of the Caddy Canyon Quartzite (as much as 2 km of orthoquartzite), Inkom Fm. (up to 150 m of olive drab to greyish red or liver-coloured siltstone) and Mutual Fm. (several hundred metres of greyish red pebbly quartzite) contains incised valley-fill conglomerates, correlated with confidence from range to range over two states (Fig. 38.2; Crittenden *et al.* 1971; Christie-Blick 1982). The succession at Pocatello is representative (Fig. 38.5).

A striking feature of the Brigham Group, in addition to its overall lithostratigraphy, is the presence at many localities of an unconformity (sequence boundary) with prominent conglomerate-filled incised valleys at or near the top of the Caddy Canyon Quartzite from Pocatello south and west to the Sheeprock Mountains (Christie-Blick *et al.* 1988; Christie-Blick & Levy 1989; Levy & Christie-Blick 1991*b*; Levy *et al.* 1994; Christie-Blick 1997). The valleys are typically as much as several tens of metres deep (up to 160 m deep in the southern Sheeprock Mountains). A second unconformity, at or near the base of the Mutual Fm., is recognized on the basis of a regionally persistent abrupt upward coarsening of facies, and the presence of incised valleys in the southern Sheeprock Mountains and central Wasatch Range (Christie-Blick 1997).

## Glaciogenic deposits and associated strata

Glaciogenic deposits in southeastern Idaho and Utah consist of a heterogeneous assemblage of sedimentary facies associations (see Link *et al.* 1994 for the most recent review) including massive diamictite, stratified diamictite and graded sandstone, diamictite and laminated mature sandstone, and carbonate, shale and sandstone. The massive diamictite association is characterized by diamictite as much as hundreds of metres thick with little or no stratification. Where present, bedding is indistinct, and defined by subtle variations in clast abundance, inverse and normal grading, and lenses and interbeds of stratified sandstone or argillite. This association is best developed in the upper Scout Mountain Member, lower part of the Mineral Fork Fm. and in the formation of Perry Canyon (Fig. 38.4). The stratified diamictite and graded sandstone association includes bedded diamictite, disorganized clast-supported conglomerate, massive to graded sandstone with parallel lamination and cross-lamination, and rhythmically bedded and laminated argillite. The association is best expressed in the lower part of the Scout Mountain Member and formation of Perry Canyon and in western locations (Sheeprock Mountains and Deep Creek Range). The diamictite and laminated mature sandstone association is characterized by stratified diamictite with contorted lenses of conglomerate and finer-grained deposits, and parallel-laminated medium- to coarse-grained sandstone. Much of the Mineral Fork Fm. and parts of the Dutch Peak Fm. and Trout Creek sequence is composed of this association. The carbonate, shale and sandstone association consists of generally laminated limestone and dolomite as much as several metres thick, laminated, cross-laminated and graded siltstone, and fine- to coarse-grained sandstone. It is present in the upper Scout Mountain Member, and the upper formation of Perry Canyon.

Striated clasts and outsized clasts in laminated facies are present widely in the massive and stratified diamictite facies associations, though they are not common. The best examples are found in the Mineral Fork Fm. of the central Wasatch Range (Christie-Blick 1983*a*), particularly at the Mineral Fork locality (striated clasts) and on the north side of Little Cottonwood Canyon (outsized clasts in laminated facies). Such isolated stones are present locally in the Dutch Peak Fm., and in the formation of Perry Canyon on Fremont Island. Striated clasts have been observed also in the Dutch Peak Fm. (Sheeprock Mountains) and in the upper diamictite of the Scout Mountain Member in Idaho.

## Boundary relations with overlying and underlying non-glacial units

The lower contacts of diamictite-bearing strata range from unconformable to concordant and perhaps conformable. Diamictite rests upon an unconformity at most exposures of the Mineral Fork Fm.

(Fig. 38.4) and locally at the base of the formation of Perry Canyon. A striated and grooved surface with roches moutonnées is preserved beneath the Mineral Fork Fm. on the north side of Big Cottonwood Canyon (Christie-Blick 1983*a*, 1997). The age of the Big Cottonwood Fm., which underlies the Mineral Fork at that locality and all others except Antelope Island, is uncertain. By correlation with the <766 Ma Uinta Mountain Group, the Big Cottonwood is perhaps no more than a few tens of millions of years older than the glacial deposits (Fig. 38.3). We know that the Big Cottonwood was sufficiently lithified to maintain a palaeogradient of <40° at valley walls without deformation, to preserve glacial grooves, and to provide a source for well-rounded clasts in the Mineral Fork Fm.

Elsewhere (Antelope Island, and locally in the northern Wasatch Range, Yonkee *et al.* 2000*b*), diamictite overlies igneous and metasedimentary rocks of Archaean to Palaeoproterozoic age. At most locations, the lower contacts of the Pocatello Fm., the formation of Perry Canyon and Sheeprock Group are not exposed. Diamictite-bearing intervals in those sections are concordant with underlying non-diamictic deposits (Fig. 38.4). The base of the Trout Creek sequence (Horse Canyon Fm.) in the Deep Creek Range is faulted.

Upper contacts with non-glacial deposits are gradational to thick siltstone or shale in the case of the Scout Mountain Member (Idaho), the Mineral Fork Fm. on Antelope Island, the formation of Perry Canyon (northern Wasatch Mountains), the Dutch Peak Fm. (Sheeprock Mountains) and the Trout Creek sequence (Deep Creek Range; Fig. 38.4). The upper contact is unconformable at most other exposures of the Mineral Fork Fm., and not exposed at Little Mountain or on Fremont Island.

A laminated dolostone (in excess of 10 m thick) overlies diamictite of the Mineral Fork Fm. on Antelope Island (Christie-Blick 1983*a*; Yonkee *et al.* 2000*a*). A correlative considerably thinner carbonate is exposed in the northern Wasatch Range, where it overlies a variety of glacial and non-glacial facies (Crittenden *et al.* 1983). A comparably thin laminated carbonate, which overlies the upper diamictite of the Scout Mountain Member east of Pocatello (Link 1983, 1987; Corsetti *et al.* 2007), is present both *in situ* and as a sedimentary breccia. The laminated dolomite at Pocatello passes up into an upward-fining succession containing several beds of limestone and an epiclastic tuff (Fanning & Link 2004, 2008; Lorentz *et al.* 2004). Laminated carbonates above diamictite are not present in the Sheeprock Mountains or in the Deep Creek Range.

## Chemostratigraphy

Cryogenian and Ediacaran strata in Idaho and Utah have been the subject of several chemostratigraphic and other geochemical and provenance studies. C-isotope data have been obtained from three relatively thin carbonate-bearing intervals in the post-glacial part of the succession (Smith *et al.* 1994; Lorentz *et al.* 2004; Corsetti *et al.* 2007), and from organic matter from the upper part of the pre-glacial Uinta Mountain Group (Dehler *et al.* 2007). Chemical Index of Alteration (CIA) data are available for the glaciogenic Mineral Fork Fm. (Young 2002), and for the Uinta Mountain Group and Big Cottonwood Fm. (Condie *et al.* 2001). Provenance has been studied widely in both pre-glacial and post-glacial deposits (Farmer & Ball 1997; Ball & Farmer 1998; Condie *et al.* 2001; Mueller *et al.* 2007).

C-isotope ratios range from −2.9 to −6.9‰ within carbonate beds in the uppermost part of the Scout Mountain Member, including the thin laminated dolostone (level 'a' and 'cap' in Fig. 38.5; Smith *et al.* 1994; Lorentz *et al.* 2004; Corsetti *et al.* 2007). The data cluster around −4.5 to −5.5‰ in two of three sections (sections 2 and 3) sampled by Lorentz *et al.* (2004), without a well-defined stratigraphic trend. Data are more scattered in a third section (their section 1). O-isotope values of −13.6 to −22.5‰ and a positive correlation with $\delta^{13}C$ in sections 1 and 2 suggest diagenetic alteration. Lorentz *et al.* (2004) nevertheless took the general consistency of C-isotope data and the absence of a diagenetic trend in the isotopic cross-plot for section 3 to indicate that measured $\delta^{13}C$ values provide a reasonable approximation for the isotopic composition of seawater at the time of deposition. C-isotope values cluster between +1.0 and −1.5‰ in the Blackrock Canyon Limestone in Idaho (level 'b' in Fig. 38.5; Corsetti *et al.* 2007). Thin layers of dolomite in the middle of the Caddy Canyon Quartzite in Idaho are characterized by $\delta^{13}C$ values between +3.9 and +8.8‰ (three measurements, level 'c' in Fig. 38.5; Smith *et al.* 1994; Corsetti *et al.* 2007).

C-isotope data have been obtained also from organic-rich shales in the Red Pine Shale in the upper part of the Cryogenian Uinta Mountain Group, stratigraphically below the glacial deposits (Dehler *et al.* 2007). Whole-rock $\delta^{13}C$ values for organic matter range from −16.9 to −30.8‰, and are comparable to values obtained from other Cryogenian marine successions, including the Chuar Group of the Grand Canyon Supergroup (Dehler *et al.* 2005*a*). Total organic carbon for the Red Pine Shale varies from 0.07 to 5.9% (Dehler *et al.* 2007).

Mudstones in the upper part of the Mineral Fork Fm. are unusually Fe-rich (*c.* 15% $Fe_2O_3$), and comparable with Fe-rich glaciogenic deposits of the Rapitan Group in the northern Canadian Cordillera (Young 2002). The carbonate-corrected CIA for diamictite in the Mineral Fork Fm. ranges from 65 to 70, values that are significantly higher than those reported for the glaciogenic Palaeoproterozoic Gowganda Fm. of Ontario, Canada (CIA < 60). The Gowganda data were taken to indicate reduced chemical weathering under frigid conditions. Data from the Mineral Fork Fm. are consistent with the incorporation of a high proportion of weathered silicates (sedimentary rocks). About 94% of clasts larger than 1 cm are sedimentary (mean of 22 counts excluding the Antelope Island locality; Christie-Blick 1983*a*). CIA values for shales and argillites in the Uinta Mountain Group and Big Cottonwood Fm. are higher than for the Mineral Fork Fm. (mostly 75–85), reflecting significant chemical weathering of their sources (Condie *et al.* 2001).

Provenance data are available for both pre-glacial and post-glacial strata. Neodymium (Nd) and Sr-isotope data from shales of the Uinta Mountain Group are consistent with a Laurentian provenance, with both Archaean and mixed Proterozoic components (Ball & Farmer 1998; Condie *et al.* 2001; Mueller *et al.* 2007). A Nd-isotope study of the Trout Creek and McCoy Creek successions, above the glacial interval, suggest a Palaeoproterozoic source within the Mojave and Yavapai provinces (Farmer & Ball 1997). Detrital zircons from Brigham Group sandstone are of Grenvillian age (1250–1000 Ma), with smaller populations of Mesoproterozoic, Palaeoproterozoic and Archaean age. These are interpreted to represent a trans-continental provenance (Stewart *et al.* 2001).

## Other characteristics

Organic-walled spheres and aggregates *c.* 5–20 mm in diameter [*Bavlinella faveolata* (Shepeleva) Vidal] have been described from the Red Pine Shale and older formations of the Uinta Mountain Group, along with locally abundant leiosphaerid acritarchs and filaments (Vidal & Ford 1985; Nagy & Porter 2005; Sprinkel & Waanders 2005; Dehler *et al.* 2007). Similar unicells, dyads and aggregates, also referred to as *Bavlinella*, were recognized by Knoll *et al.* (1981) in mudrocks of the Mineral Fork Fm. In some cases, organic material appears to have been pyritized or to be composed of abiotic framboidal pyrite (Dehler *et al.* 2007; N. J. Butterfield, pers. comm. 2009). However, the existence of at least some microfossils is supported by a restricted stratigraphic range, the presence of transitional morphologies, and in the case of the Mineral Fork, the absence of pyrite (Knoll *et al.* 1981; S. Porter, pers. comm. 2009). Trace fossils, including *Skolithos*, are present widely in the upper part of the Brigham Group (Camelback Mountain Quartzite and correlatives; Fig. 38.4), consistent with a Cambrian age.

## Palaeolatitude and palaeogeography

Hematite-cemented sedimentary rocks in the Uinta Mountain Group yield a well-defined palaeomagnetic pole with a mean of 0.8°N, 161.3°E ($a_{95} = 4.6°$; $n = 9$ sampling localities consisting of 79 sites; Weil *et al.* 2006). The characteristic remanent magnetization (ChRM) is east-directed (or antipode), of low positive or negative inclination, and typically unblocked over a narrow range of high laboratory temperatures between 660 °C and 680 °C. A second magnetization is north- to NE-directed with moderate to steep inclination. The presence of dual polarities suggests that the ChRM was acquired close to the time of deposition, and over a span long enough to average out secular variation of the geomagnetic field. The second magnetization is inferred to be a recent or modern overprint. The data are consistent with a low palaeolatitude for Laurentia during deposition of Uinta Mountain Group sediment, with a tight counterclockwise apparent polar wander path in the south Pacific, and with deposition over an interval that was sufficiently short (tens of millions of years) for the palaeolatitude not to have changed significantly during deposition. Sampling for this most recent study was more comprehensive than that undertaken by Bressler (1981), but inferences about pole location and palaeolatitude are not statistically distinguishable.

Palaeomagnetic data are not available for the glacial interval or for the later Cryogenian and Ediacaran in the area of interest. Suitable rocks are not present in the glaciogenic section, which in most places is sufficiently deformed and metamorphosed (greenschist) to be problematic. Nor has it been possible to achieve a positive fold test, though that was tried twice in the Brigham Group near Huntsville, Utah (as reported in Link *et al.* 1994). Based on a global assessment of the reliability of the best available palaeomagnetic data from Cryogenian strata and interpreted positions of continents, Evans *et al.* (2000) concluded that glacial deposits in southeastern Idaho and Utah accumulated at a palaeolatitude of $<5°$, and that western North America remained at low palaeolatitude through the end of the Ediacaran (cf. Torsvik *et al.* 1996).

## Geochronological constraints

A population of detrital zircons from the formation of Outlaw Trail, in the lower part of the eastern Uinta Mountain Group, yields a concordia age of 766 ± 5 Ma, suggesting the rocks must be younger (Dehler *et al.* 2007). This is substantially younger than previous estimates of *c.* 900 Ma for the upper part of the Uinta Mountain Group (Link *et al.* 1993; Stone 1993), and it places a very conservative upper bound on the age of pre-glacial sedimentation in Utah. Based on similarities in microfossils to Chuar Group strata in the Grand Canyon that contain a 740 Ma tuff bed, the Uinta Mountain Group is estimated to span 766–740 Ma, and to be Cryogenian in age. The Big Cottonwood Fm. is estimated to span the same time interval.

The timing of Neoproterozoic glaciation in the Pocatello Fm. of southeastern Idaho is bracketed by SHRIMP U–Pb dating of volcanic clasts within diamictite and two intervals of tuff in the Scout Mountain Member. Felsic volcanic clasts in diamictite near Pocatello are dated as 717 ± 4 Ma (Fanning & Link 2004) and 701 ± 4 Ma (Fanning & Link 2008). These cobble- to boulder-sized clasts are interpreted as having been eroded from uplifted exposures of the underlying Bannock Volcanic Member of the Pocatello Fm. (Link 1983). On Oxford Mountain near the Idaho–Utah state line (Fig. 38.1), an epiclastic plagioclase-phyric mafic tuff breccia interbedded with diamictite-bearing rocks near the base of the Scout Mountain Member yielded a SHRIMP U–Pb concordia age of 686 ± 4 Ma (Fanning & Link 2008; a separate older population of zircons (709 ± 5 Ma) was reported by Fanning & Link 2004). Much or all of the Scout Mountain Member is thus younger than 686 Ma. Above the upper Scout Mountain diamictite and the overlying laminated carbonate, an epiclastic tuff contains zircons dated as 667 ± 5 Ma (Fanning & Link 2004).

The more stratigraphically complete sections of northern and west-central Utah remain to be dated. Attempts to separate zircons from a clast of rhyolite obtained from conglomerate near the base of the Dutch Peak Fm. (Sheeprock Mountains) and from a suite of intermediate-composition volcanic clasts collected from the formation of Perry Canyon at Little Mountain were not successful.

Cryogenian magmatism is also recognized north of the Snake River Plain in central Idaho. Three samples of felsic orthogneiss exposed in Wildhorse Creek of the Pioneer Mountains yield SHRIMP U–Pb zircon upper-concordia intercept ages of 692.3 ± 5.2, 695.7 ± 8.0 and 696.5 ± 9.0 Ma (location in Fig. 38.1; K.M. Durk-Autenrieth, Idaho State University, pers. comm. 2007). The age is thus close to 695 Ma. A somewhat older U–Pb age (725 ± 5 Ma) was obtained for the House Mountain orthogneiss near the southern edge of the Atlanta lobe of the Idaho batholith (location shown on Fig. 38.l; M. Schmitz, Boise State University, pers. comm. 2006). In central western Idaho, 684 Ma volcanic rocks overlie diamictite near Edwardsburg (Lund *et al.* 2003).

Lund *et al.* (2010) report ages of 665–650 Ma for Cryogenian alkalic plutonic rocks of the Beaverhead–Big Creek belt in central Idaho. As these are intrusive rocks, within exposures of Proterozoic basement, their relations to Neoproterozoic glaciogenic successions are not clear. However, these new ages extend the locations of Cryogenian magmatism into a NW-trending swath across much of Idaho.

The only other direct age constraint on the Neoproterozoic of Utah and Idaho is a 580 ± 7 Ma age (Fig. 38.2; $^{40}Ar/^{39}Ar$ on hornblende from trachyte from the upper Browns Hole Fm.; Christie-Blick & Levy 1989). This unit overlies strata that contain incised valleys interpreted to represent sea-level drawdown associated with a younger Cryogenian glaciation (Christie-Blick & Levy 1989).

## Discussion

The Neoproterozoic and early Cambrian palaeoenvironmental and palaeogeographic evolution of Idaho and Utah reflect a combination of climatic and tectonic controls. Mostly marine glacial deposits are dated in southeastern Idaho as *c.* 686 ± 4 Ma to *c.* 667 ± 5 Ma (Cryogenian). Thicker, more complete sections in northern and west-central Utah may include older glacial deposits, but age data are not yet available for those rocks. Direct evidence for a younger Cryogenian glaciation is not preserved in this region. However, stratigraphically restricted incised valleys as much as 160 m deep are inferred to be at least in part of glacial-eustatic origin. Available evidence suggests that all of the Neoproterozoic deposits accumulated at low palaeolatitude.

### *Depositional settings and climatic controls*

The Uinta Mountain Group accumulated in braided fluvial and shallow marine environments (Wallace & Crittenden 1969; Crittenden & Wallace 1973; Sanderson 1984; Dehler *et al.* 2005*b*, 2007, 2010). Beginning at the base, marine intervals have been documented in the Jesse Ewing Canyon Fm. and in the informal formations of Red Castle, Mount Agassiz, Dead Horse Pass, Moosehorn Lake and Red Pine Shale (Fig. 38.3; Dehler *et al.* 2007; E. M. Kingsbury, pers. comm. 2009). Tidal rhythmites preserved in the coeval Big Cottonwood Fm. provide some of the best evidence for a marine connection in spite of an intracratonic setting that must have been many hundreds of kilometres from the nearest oceanic crust (Chan *et al.* 1994; Ehlers *et al.* 1997; Ehlers & Chan 1999; Dehler *et al.* 2005*b*).

Glaciogenic and associated deposits in southeastern Idaho and Utah accumulated for the most part in a deep marine setting (water depths as great as hundreds of metres; Christie-Blick 1983*a*, 1997; Crittenden *et al.* 1983; Link 1983; Link *et al.* 1994). This is indicated by the overall character of the facies, by their association with turbidite sandstones and subaqueous extrusive volcanic rocks, by the presence of ice-rafted dropstones and till clots (albeit sporadically), and in the Sheeprock Mountains by a progradational stratigraphic architecture. The Mineral Fork Fm. is thought to have accumulated close to the grounding line of a partially buoyant ice sheet, on the basis of glacial grooves and probable roches moutonnées at the basal contact, syndepositional deformation, and abundant well-stratified sandstone interpreted as subaqueous outwash. Direct evidence for grounding is not present elsewhere, although quartzite in the Otts Canyon and Dutch Peak formations of the Sheeprock Mountains may have accumulated in a glacial–fluvial to braid-delta setting (Link *et al.* 1994). The appreciably greater water depths implied by most of the deposits, in comparison with 'syn-rift' strata at lower and higher stratigraphic levels, is thought to indicate an isostatic response to the weight of the nearby ice sheet as well as tectonically driven subsidence.

Up to several hundreds of metres of siltstone that in most places directly overlie glaciogenic deposits, locally with multiple beds of intervening carbonate, are attributed to glacial–eustatic sea-level rise combined with thermally driven tectonic subsidence. C-isotope data from the thin laminated carbonate beds that overlie Scout Mountain Member diamictite are comparable with the strongly depleted $\delta^{13}C$ values of cap carbonates elsewhere (Corsetti *et al.* 2007). Near-zero to positive $\delta^{13}C$ signatures in stratigraphically higher carbonate layers are comparable with data from the Johnnie Fm. of Death Valley (Corsetti & Kaufman 2003) and are consistent with a late Cryogenian age.

The Brigham Group and correlatives are dominated by braided fluvial, braid delta and shallow marine deposits (Link *et al.* 1987; Christie-Blick & Levy 1989; Levy & Christie-Blick 1991*b*; Levy *et al.* 1994; Christie-Blick 1997). Incised valleys at several scales and horizons, at or near the top of the Caddy Canyon Quartzite, represent approximately the same stratigraphic level as the glaciogenic Ice Brook Fm. in the Mackenzie Mountains of northwestern Canada (Aitken 1991; Levy & Christie-Blick 1991*b*; Levy *et al.* 1994; Ross *et al.* 1995; Christie-Blick 1997). Taken together, these observations are consistent with a glacial–eustatic origin and a late Cryogenian age (655–635 Ma; Hoffman & Li 2009).

The origin of the sequence boundary at or near the base of the Mutual Fm. is less clear (Levy & Christie-Blick 1991*b*; Christie-Blick 1997). Eustatic and tectonic explanations are both permitted by the only available age constraint (an $^{40}Ar/^{39}Ar$ age of $580 \pm 7$ Ma from the overlying Browns Hole Fm. at Huntsville, Utah). A glacial–eustatic origin is consistent with age estimates for the Gaskiers glaciation of Newfoundland. (ID-TIMS dates on ash beds below, within and above the glaciogenic Gaskiers Fm. constrain its age to *c.* 584–582 Ma; Bowring *et al.* 2003.) A difficulty with the glacial–eustatic hypothesis is that while sea-level rise after an initial drawdown accounts for the thickness of the Mutual (hundreds of metres) over a broad area, it does not explain why fluvial sedimentation continued. Christie-Blick (1997) suggested that this might have to do with an increase in sediment supply. A difficulty with a tectonic interpretation for the unconformity is that regional uplift reduces available sedimentary accommodation. So renewed subsidence is then required to account for the considerable thickness of overlying quartzite.

### *Tectonic setting*

The rocks of southeastern Idaho and Utah are inferred to have accumulated in rift-related basins associated with the development of a passive continental margin, remnants of which are preserved today from eastern Alaska to eastern California (Stewart 1972; Stewart & Suczek 1977; Bond *et al.* 1983, 1985; Christie-Blick & Levy 1989; Ross 1991; Levy & Christie-Blick 1991*a*). Details of the tectonic history are unresolved, particularly with respect to the timing of continental separation. Evidence for crustal extension is necessarily indirect owing to the difficulty in documenting stratigraphic growth in available outcrop and to the tendency for originally faulted basin margins to be offset by younger structures. The pre-glacial Uinta Mountain Group and Big Cottonwood Fm. are thought to have been deposited in an east-trending <766 Ma rift basin, based upon abrupt thinning of these deposits towards the north across a fault-controlled boundary (Sears *et al.* 1982; Christie-Blick 1997; Mueller *et al.* 2007). Stone's (1993) alternative interpretation of the Uinta Mountain Group in terms of a seaway that encroached on an area of low topography fails to account for a stratigraphic thickness as great as 7 km. Abrupt thickness changes in the Sheeprock Group of the Sheeprock Mountains are attributed to a combination of facies change, progradation into a deep depocentre, and tectonic tilting towards the SE at a hinged basin margin (Christie-Blick 1982, 1997; Crittenden *et al.* 1983). Inferred stratigraphic growth encompasses the interval of glaciation, although it may have begun earlier. Magmatism, assumed to be rift-related, is widespread at this same stratigraphic level, ranging in age from *c.* 720 to *c.* 685 Ma, and as young as 650 Ma in the case of intrusive rocks in central Idaho. The Bannock Volcanic Member of the Pocatello Fm. includes mafic volcanic rocks with a rift-related trace element signature (Harper & Link 1986).

Facies and thickness variation at the level of the upper member of the Pocatello Fm. and Brigham Group (post-glacial) are less pronounced, consistent with the onset of thermally driven subsidence of a passive margin as early as *c.* 665 Ma. However, quantitative analysis of tectonic subsidence in Cambro-Ordovician strata indicates that a second pulse of thermal subsidence began as late as *c.* 520 Ma, if account is taken of recent changes to the Cambrian timescale (Bond *et al.* 1983; Christie-Blick & Levy 1989; Levy & Christie-Blick 1991*a*; Link *et al.* 1994). Early Cambrian timing for this second pulse is supported in the central Wasatch Range by angular discordance of up to 10° between the Big Cottonwood Fm. and overlying Cambrian quartzite, which Christie-Blick & Levy (1989) and Christie-Blick (1997) took to indicate rift-related reactivation of the northern bounding fault of the Big Cottonwood basin. The mismatch between the magnitude of early Palaeozoic post-rift thermal subsidence and the minimal evidence for crustal extension after *c.* 665 Ma is best explained in terms of inhomogeneous extension of the lithosphere in latest Neoproterozoic time (Christie-Blick & Levy 1989). Stratigraphic evidence for multiple rifting events and for one or more times of passive margin formation at or after *c.* 665 Ma is inconsistent with palaeomagnetically based interpretations of a rift-to-drift transition prior to 750 Ma (Wingate & Giddings 2000; Torsvik 2003; Weil *et al.* 2006). The latter requires assumptions about the identity of counterpart blocks, and about the positioning of those blocks at the time of continental break-up.

### *Regional correlations*

Available stratigraphic and geochronological data support regional correlation. The pre-glacial Uinta Mountain Group and Big Cottonwood Fm. are comparable in terms of age and tectonic setting to the Chuar Group of the Grand Canyon region (Karlstrom *et al.* 2000). Marine strata within the three successions are approximately coeval (*c.* 766–740 Ma), suggesting that they may have accumulated within the same epicontinental 'ChUMP seaway', which Dehler *et al.* (2005*b*, 2007) regarded as encompassing at least part of the Pahrump Group in the Death Valley area of eastern California. Cryogenian glacial deposits of Idaho and Utah are inferred to correlate approximately with the Kingston

Peak Fm. (upper Pahrump Group; Stewart & Suczek 1977; Link *et al.* 1993; Corsetti *et al.* 2007; Mrofka & Kennedy 2011; Petterson *et al.* 2011) and the Edwardsburg Fm. in central Idaho (Lund *et al.* 2003, 2011), although a glacial origin for the diamictite-bearing Edwardsburg Fm. has not been demonstrated. The successions in southeastern Idaho and Utah have also traditionally been broadly correlated with the Windermere Supergroup in Canada (Link *et al.* 1993; Lund *et al.* 2003), although age control is not yet adequate to explore diachrony in the timing of either glaciation or tectonism in western North America, as predicted by the 'zipper-rift' hypothesis of Eyles & Januszczak (2004).

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