Sedimentary record of seismic events in the Eocene Green River Formation and their implications for regional tectonics on lake evolution (Bridger Basin, Wyoming)

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Abstract

Outcrops and cores from the top of the lacustrine Tipton Member and the base of the Wilkins Peak Member (~51.5 Ma) of the Eocene Green River Formation, Bridger Basin in southwestern Wyoming yield a wide variety of sedimentary deformation features many of which are laterally extensive for more than 50 km. They include various types of folds, load structures, pinch-and-swell structures, microfaults, breccias and sedimentary dikes. In most cases deformation is represented by hybrid brittle–ductile structures exhibiting lateral variation in deformation style. These occur in low-energy, profundal organic-rich carbonate mudstones (oil shales), trona beds, tuffs, and profundal to sublittoral silty carbonate deposited in paleolake Gosiute. The deformation is not specific to the depositional environment because sedimentary units stratigraphically higher with similar facies show no deformation.

The studied interval lacks any evidence for possible trigger mechanisms intrinsic to the depositional environment, such as strong wave action, rapid sediment loading, evaporite dissolution and collapse, or desiccation, so ‘endogenic’ causes are ruled out. Thus, the deformation features are interpreted as seismites, and change in deformation style and inferred increase in intensity towards the south suggest that the earthquakes were sourced from the

nearby Uinta Fault System. The 22 levels exhibiting seismites recognized in cores indicate earthquakes with minimum magnitudes between 6 and 7, minimum epicentral intensity (MCS) of ~9, and varying recurrence intervals in the seismic history of the Uinta Fault System, with a mean apparent recurrence period of ~8.1 k.y. using average sedimentation rates and dated tuffs; in detail, however, there are two noticeably active periods followed by relative quiescence. The stratigraphic position of these deformed intervals also marks the transition between two distinct stages in lake evolution, from the balanced-filled Tipton Member to the overlying, underfilled Wilkins Peak Member. Thus, these seismites are evidence for regional-scale changes in lacustrine sedimentation of Eocene Lake Gosiute in response to syndepositional tectonic activity. Analysis of synsedimentary deformation features is, therefore, a promising yet under-utilized tool to trace the tectonic evolution of lacustrine deposits of the Green River Formation and other tectonically active marine and non-marine basins.

Keywords: Green River Formation, seismites, lacustrine facies, oil shales, syndepositional tectonism, Eocene

1. Introduction

Lacustrine sediments provide an unparalleled opportunity to study the sedimentary record of ancient earthquakes in continental basins. Lake deposits typically consist of rheologically susceptible, heterolithic sediments deposited in an overall low-energy setting (Renaut and Gierlowski-Kordesch, 2010), which increases the potential for preservation of deformation features and eliminates aseismic trigger mechanisms specific to other depositional environments (e.g., Sims, 1975; Moretti and Sabato, 2007). The sedimentary record of lacustrine basins has therefore been used successfully in paleoseismological studies of Quaternary deposits (e.g., Doig, 1986, 1990; Marco et al., 1996; Chapron et al., 1999; Rodríguez-Pascua et al., 2010; Beck, 2011; Strasser et al., 2013), and to identify tectonically

active phases in older successions (e.g., Moretti and Sabato, 2007; Berra and Felletti, 2011; Törő and Pratt, 2015a, b). Embracing the stratigraphic record of sedimentary deformation as a valid tectonic archive in low-energy lacustrine and marine strata opens the door to new avenues of appreciation of such basins and the rheological attributes of their sediments (e.g., Pratt, 1998b, 2001; Martín-Chivelet et al., 2011; El Taki and Pratt, 2012; Törő and Pratt, 2015a).

The Green River Formation (Eocene) represents one of the best documented ancient lake systems worldwide (e.g., Smith et al., 2008; Smith and Carroll, 2015) and lacustrine deposits of the Greater Green River Basin in southwestern Wyoming are type examples of evolutionary lake phases (Carroll and Bohacs, 1999; Bohacs et al., 2000). This lake system was influenced by syndepositional tectonic activity related to the Sevier Fold and Thrust Belt and Laramide orogeny (Love, 1970; Steidtmann et al., 1983; Dickinson et al., 1988; DeCelles, 1994; Carroll et al., 2006; Smith et al., 2008). Although the sedimentology of the Green River Formation in the Greater Green River Basin or in adjacent Colorado and Utah has been studied extensively, only a few have recorded the presence of sedimentary deformation features and these have done so only in passing (e.g., Bradley, 1930; Picard, 1966; Dyni and Hawkins, 1981; Smoot, 1983). Sedimentary deformation features therefore comprise an essentially unexplored aspect of the geology of the Green River Formation (Törő et al., 2015; Törő and Pratt, 2015a, b).

Törő et al. (2015) reported sedimentary deformation features from the lowermost Wilkins Peak Member of the Green River Formation and discussed syndepositional tectonics in relation to the regional basin evolution at the time (~51.5 Ma). That study was limited to outcrop data in the Bridger Basin of southwestern Wyoming, where sedimentary deformation features had not been reported previously. Here, we expand on this by describing the full suite of these kinds of features preserved in the uppermost ~10 m of the balanced-fill Tipton Member and the lowermost underfilled Wilkins Peak Member by combining observations both in cores and of outcrops over a larger area.

We studied the origin of deformed horizons using facies analysis of the host lacustrine deposits, and the stratigraphic, macroscopic, and microscopic attributes of the deformation features. This allowed us to: (1) determine the timing of deformation relative to shallow burial; (2) interpret the rheological properties of the sediments at that time; (3) deduce the most likely trigger mechanism; and ultimately (4) present the sedimentary record of late-stage Laramide tectonic movements that affected the evolution of Eocene Lake Gosiute. Our study shows that understanding the deformation features at certain stratigraphic horizons and the physical conditions under which these features formed can enhance interpretations of the evolution of the paleolake basin, and enable the refinement of the paleoenvironmental setting and the paleotectonic history of the region.

2. Geological and stratigraphic framework

2.1. Regional geology

Lacustrine sediments of the Green River Formation were deposited in three interconnected paleolakes in the foreland of the Sevier Fold and Thrust Belt of the central Rocky Mountain region: (1) Fossil Lake (Fossil Basin, southwestern Wyoming); (2) Lake Gosiute (Greater Green River Basin, southwestern Wyoming and adjacent Utah and Colorado); and (3) Lake Uinta (Piceance Creek Basin, northwestern Colorado, and Uinta Basin, northeastern Utah) during the early to middle Eocene (53–43 Ma) (Smith et al., 2008 and references therein) (Fig. 1).

The Greater Green River Basin is bounded by fault zones related to Laramide-style uplifts to the north (Wind River and Gros Ventre Mountain thrusts), east (Granite Mountain and Sierra Madre thrust), and south (Axial Basin Arch and the fault system of the Uinta Uplift), and the Sevier Fold and Thrust Belt to the west (Witkind and Grose, 1972; Love, 1970; Steidtmann et al., 1983; Dickinson et al., 1988; Steidtmann and Middleton, 1991; Roehler, 1992b, 1993; DeCelles, 1994, 2004; Johnston and Yin, 2001; Chamberlain et al., 2003; Yonkee and Weil, 2010). Intrabasinal deformation during the Laramide orogeny formed anticlinal structures (e.g., Rock Springs Uplift, Cherokee Ridge, Wamsutter Arch) that

subdivided the basin into four subbasins: the Bridger (or Green River), Great Divide, Washakie, and Sand Wash basins (Love et al., 1961; Mederos et al., 2005; Bader, 2008; Johnson and Anderson, 2009). These structures were intermittently active during deposition of the Green River Formation, and are assumed to have influenced the paleogeomorphology, basin shape, gradients, the type and distribution of depositional environments, as well as the salinity of Lake Gosiute (e.g., Bradley, 1964; Bradley and Eugster, 1969; Surdam and Stanley, 1979, 1980; Roehler, 1968, 1992b; Pietras et al., 2003; Pietras and Carroll, 2006; Carroll et al., 2006, 2008; Smith et al., 2008; Doebbert et al., 2010). Short-term, climatically driven changes have been also recognized on the scale of lithologic cycles (Bradley, 1929; Fischer and Roberts, 1991; Roehler, 1993; Birgenheier and Vanden Berg, 2011; Aswasereelert et al., 2013). However, direct correlation between climatic changes and large-scale, basin-wide evolution remains problematic (Smith et al., 2008).

In general, coarse clastic and fluvial sediments belonging to the Wasatch Formation were deposited near to the uplifted edges of the basin, while lacustrine mudstones, carbonates and evaporites of the Green River Formation were deposited towards the basin centre (Roehler, 1992b, 1993). These lacustrine deposits are the classic examples of lake-basin types described by Carroll and Bohacs (1999), and are subdivided into four members (Roehler, 1991a, b): the Luman Tongue, Tipton Member, Wilkins Peak Member, and Laney Member (Fig. 2A). The latter three members record the dynamic evolution of Lake Gosiute from paleohydrologically open to closed and then back to an open state (Surdam and Stanley, 1980; Carroll and Bohacs, 1999; Bohacs et al., 2000; Smith et al., 2008). The depositional history, stratigraphy, and geochemistry of the Wilkins Peak Member are well documented (Eugster and Surdam, 1973; Eugster and Hardie, 1975; Smoot, 1983; Roehler, 1992b; Pietras et al., 2003; Pietras and Carroll, 2006; Aswasereelert et al., 2013; Murphy et al., 2014; Smith et al., 2014a, 2015). Studies of the Tipton Member, however, have been mostly at a local scale or as part of regional studies (Pipiringos, 1955; Roehler, 1968, 1990, 1992a, b; Graf et al., 2015), and consequently there is little information on the sedimentology of the transition with the overlying Wilkins Peak Member.

Despite numerous multidisciplinary studies, certain aspects of the Green River Formation remain unexplored, including the origin of various sedimentary deformation features associated with both profundal deposits (e.g., oil shales) and marginal facies (e.g., cross-laminated or -bedded calcareous sandstones and siltstones). These deformation structures are important because many have been used as evidence of depositional setting (e.g., Bradley, 1973; Dyni and Hawkins, 1981; Smoot, 1983), and potential misinterpretations may have led to an inaccurate picture of the evolution of the Green River system.

This study focuses on sedimentary deformation structures at the boundary of the Tipton and the Wilkins Peak members (~51.5 Ma; Smith et al., 2008, 2010) west of the Rock Springs Uplift (Figs. 1, 2A, B). The boundary records a major change in the evolution of paleolake Gosiute, where lake conditions are interpreted to have changed from a fluctuating, balanced-filled lake to an underfilled, evaporative, playa-type lake (Carroll and Bohacs, 1999).

2.2. The Tipton–Wilkins Peak boundary

In outcrop studies the boundary of the two members has been placed at the top of a laterally extensive stromatolite bed in the northeastern part of the Bridger Basin or at a sharp change in the colour and lithology from thick, dark-brown or dark-grey oil shales to light-grey or light-green mudstone and siltstone in the east-central area of the basin (Bradley, 1964; Roehler, 1989, 1991b, 1992a, b; Pietras et al., 2003; Walker, 2008; Graf et al., 2015). Farther to the south, Roehler (1989) placed the top of the Tipton Member at a laterally extensive tuff near Little Mountain and Red Creek, while Smith et al. (2014a) used a stromatolite bed near Currant Creek and a sandstone bed at Little Mountain (Fig. 2B). In cores the boundary is usually placed where there is a sharp reduction in Fischer Assay oil-yield quantities, which also coincides with an abrupt change in colour and lithology, along with the first appearance of thin trona beds in the inferred area of the depocentre (e.g., Dana and Smith, 1972; Roehler, 1991a; Smith et al., 2010, 2014a; Self et al., 2011).

Most published descriptions and stratigraphic sections from the studied interval do not record synsedimentary deformation features. There are, however, a few exceptions: Bradley (1964) described ‘crudely bedded’ sediments, ‘mudcracks’, ‘subaqueous deformation’ and ‘wrinkled claystone’ at the boundary; Roehler (1968) mentioned unusual ‘mudcracks’ in the southeastern part of the basin that are not present elsewhere; Walker (2008) and Smith et al. (2014a) described rip-up clasts, breccias, convolution and ‘mudcracks’ from the transitional interval. Nonetheless, in all cases detailed descriptions of these features are lacking.

Eight localities were chosen to study the transitional interval in the southeastern part of the Bridger Basin, near the depocentre (Figs. 1, 2B). Five outcrop sections were measured at centimetre- to decimetre-scale along the Flaming Gorge Reservoir including, from north to south, Slippery Jim Canyon (SJC), Middle Firehole Canyon (MFC), Firehole Canyon (FC), Sage Creek (SC), and Red Creek (RC) (Figs. 2B, 3). For the MFC section, deformation features were studied at three locations, and at several more places in road cuts between Wilkins Peak and SJC (Fig. 2B). To supplement field data three cores were described: the ERDA White Mountain No. 1 (WM1), ERDA Black Forks No. 1 (BF1), and DOE/LETC Currant Creek Ridge No. 1 (CCR1) (Figs. 2B, 4), located ~10–15 km east–northeast of the line of the outcrop sections. Because the uppermost Tipton Member at the study area usually crops out as vertical cliffs or slopes mostly covered with debris (Fig. 2C), observations on this unit were limited almost exclusively to cores.

The Tipton–Wilkins Peak boundary in the northern part of the study area (all locations except RC) is marked by a gradational and conformable transition from black to dark-brown, grey to bluish-weathering, organic-rich dolomicritic mudstone (oil shales; Tipton Member) into grey and greenish-grey, grey- to orange-weathering, organic-poor mudstone and siltstone (Wilkins Peak Member) over a 1–2 m interval (Figs. 3, 4). This transition into the basal Wilkins Peak is a conspicuous, bench-forming lithological unit in the southeastern Bridger Basin (Fig. 2C) and it is succeeded by several 3–5 m thick, coarsening-upward successions.

Along White Mountain (Fig. 2B), the boundary becomes recessive as a result of a slight change in facies from siltstone to mudstone. Therefore outcrops have not been studied in detail north of Interstate 80. Toward the south, at RC, the transition is not as sharp and occurs over a ~25 m thick interval where the lithologies of the two members interfinger. The tuff bed used by Roehler (1989) was not found, so the base of the Wilkins Peak Member was placed at the first appearance of extensive sedimentary dikes, which coincides with the disappearance of organic-rich mudstone intervals and the beginning of dominance of grey carbonate mudstone interbedded with calcareous siltstone and sandstone (Fig. 3).

The depocentre, and probably the deepest part of the lake, was located about 5 km south of the CCR1 drillsite during the deposition of the Tipton Member (Bradley, 1964; Roehler, 1992b) (Fig. 1). Later, during deposition of the Wilkins Peak Member, it shifted towards the east–northeast, based on thicknesses and the location of trona-bearing intervals (e.g., Culbertson, 1971; Roehler, 1991a, 1992b). Consequently, all the studied sections represent the deeper, central part of Lake Gosiute, except RC, which was probably located on a more elevated area at the southern flank of the Rock Springs Uplift, and somewhat closer to the paleolake margin (Bradley, 1964; Roehler, 1992b).

2.3. Sedimentary facies

2.3.1. Description

The uppermost Tipton Member has a relatively uniform lithology, dominated by finely laminated carbonate mudstone with variable organic content (Figs. 3, 4), plus interbedded thin tuffs and silty mudstone that are massive or locally upward-fining. Scattered casts of evaporite crystals, up to 1 cm in diameter, may be present locally within the laminated mudstone close to the Tipton–Wilkins Peak boundary. In the studied sections there is no sedimentary evidence to suggest subaerial conditions, such as desiccation cracks, bird tracks, or root traces.

The lowermost Wilkins Peak Member is characterized by parallel-laminated carbonate mudstone, with interlaminated planar- or lenticular-bedded calcareous siltstone and subordinate sandstone (Figs. 3, 4). These lithologies are organized into coarsening-upward cycles characterized by a gradual upward decrease in organic content of the mudstone, as represented by change in colour from brown or dark-grey to light-grey, and increase in thickness of laminae with a concomitant disappearance of distinct lamination. Fine- to medium-grained calcareous sandstone lenses, laminae and thin beds are locally interbedded towards the top of each cycle. Sandstones display parallel or wavy lamination, wave-ripple cross-lamination, or, less commonly, combined-ripple cross-lamination. The cycles show an overall coarsening trend from the northernmost outcrop, SJC, south to SC, with an increasing amount of interbedded very fine- to medium-grained sandstone. Similar to the Tipton Member, the studied interval lacks ostracode or ooid grainstone interbeds, and no trace or body fossils (e.g., fish, gastropods, bivalves, wood debris, leaves) are present. At RC two 2–3 cm thick microbial carbonate beds are characterized by irregularly wavy parallel lamination. The stratigraphically lower bed occurs in a uniform, planar-laminated, organic-poor carbonate mudstone interval, several metres thick, whereas the upper bed is associated with a transition from parallel-laminated carbonate mudstone to wavy, lenticular laminated calcareous siltstone (Fig. 3). The three cores represent a somewhat deeper water facies, and the Wilkins Peak Member is finer grained, with few interbedded sandstones, and the coarsening-upward trends become less noticeable. Bedded trona occurs in the BF1 core (Roehler, 1991a, c; Smith et al., 2010, 2014a) (Fig. 4).

2.3.2. Interpretation

Finely laminated oil shales of the Tipton Member indicate deposition from suspension below storm wave base (e.g., Roehler, 1990; Walker, 2008; Tänavsuu-Milkeviciene and Sarg, 2012). High organic content shows that sedimentation took place under low-oxygenated or anoxic conditions in a stratified lake (e.g., Demaison and Moore, 1980). Small trona casts in the laminated oil shales likely

formed intrastratally as displacive crystals from saturated brines (Pietras and Carroll, 2006; Jagniecki and Lowenstein, 2015).

The lowermost cycles of the Wilkins Peak Member are different from those higher in the succession which exhibit playa indicators such as evaporite mineral casts, polygonal desiccation features, and subaerial and pedogenic modifications, including calcareous nodules, root traces, bioturbation, and evaporite mineral-filled pores (e.g., Eugster and Surdam, 1973; Smoot, 1983; Pietras and Carroll, 2006). Rather, they are more akin to the Tipton Member in lacking such features (e.g., Roehler, 1968, 1990; Walker, 2008; Graf et al., 2015). Thus, they are interpreted to have been deposited in the shallow profundal to sublittoral or “offshore zone” (Roehler, 1990) below storm wave base or between storm wave base and fair-weather wave base (Tānavsū-Milkeviciene and Sarg, 2012). The lack of fossils and trace fossils points to the likelihood of saline conditions. Ripple cross-laminated sandstones indicate deposition in shallower sublittoral zones along wave-agitated lake margins as affected by short-lived storm events (Tānavsū-Milkeviciene and Sarg, 2012). Microbial carbonate beds in the Green River Formation are usually interpreted as having formed in the littoral zone of the lake margin (Wolfbauer and Surdam, 1974; Roehler, 1992a, b; Rhodes et al., 2002). The microbial beds at RC, however, differ because they are not associated with sediments indicating wave agitation, such as ooid or pisoid grainstone. Thus, they are interpreted as the products of deeper water, sublittoral microbial mats (Sarg et al., 2013). The lowermost Wilkins Peak cycles show an overall shallowing upwards aspect, which suggests gradual shrinking and shallowing of Lake Gosiute, with the contemporary paleoshoreline located in the western Washakie Basin ~80 km east of the outcrops (Roehler, 1992b, 1993). Bedded trona is thought to have formed mainly in the depocentre during desiccating, hypersaline playa-type lake phases (e.g., Pietras and Carroll, 2006; Jagniecki and Lowenstein, 2015). Their microstructural characteristics suggest formation under water, within the unlithified sediments, or diagenetically after burial (e.g., Jagniecki and Lowenstein, 2015). Efflorescent salt crusts of trona may have also precipitated on the subaerially exposed mudflats along the lake margins (Deardorff and Mannion, 1971; Eugster and Hardie, 1975; Smoot, 1983). The trona deposits in BF1 formed synchronously with profundal to

sublittoral deposits, which suggests that they must have formed in a deeper water setting from saturated bottom brines in a stratified lake (e.g., Jagniecki and Lowenstein, 2015).

3. Sedimentary deformation structures

Deformation features in the Tipton Member have been observed mainly in cores, where they form up to twelve distinct horizons hosted by laminated oil shale (Fig. 4). In outcrops and cores of the Wilkins Peak Member at the northern study area (all locations except RC) sedimentary deformation structures are abundant only in lowest two coarsening-upward cycles (Figs. 3, 4). These cycles are well exposed south of the town of Green River where they form an orange-weathering unit ~7 m thick that extends throughout more than 20 km laterally (Figs. 2B, C). They host five to ten deformed layers that are separated from each other by undeformed or slightly deformed intervals showing gradual increase in intensity of deformation towards the deformed layer. The style of deformation is fairly uniform throughout the northern area and represented by a large variety of millimetre- to centimetre-scale hybrid brittle–ductile deformation features. Lacustrine sediments at RC, however, show almost exclusively sedimentary dikes in fourteen deformed beds that developed in a ~25 m thick interval comprising five sedimentary cycles. The overlying coarsening-upward cycles are identical in lithology, but lack deformation features at all localities studied.

3.1. Ductile Deformation Structures

3.1.1. Folds and convolute lamination

3.1.1.1. Description

Folds occur in interbedded laminated oil shale and massive silty mudstone of the Tipton Member (Figs. 5A–C). In the Wilkins Peak Member folding developed in calcareous siltstone interbedded with carbonate mudstone or sandstone, which gives a distinct crumpled appearance (Figs. 5D–F, C).

Individual folds are typically irregularly spaced 1–25 cm apart and mostly 2 or 3 cm in size, although some reach up to 15 cm in height and 10 cm in width. Fold geometry is variable, with open to tight and recumbent shapes. There is no particular orientation of the fold axes, not even within the same interval. Locally, the deformed beds are convoluted, showing irregular folding, with variable amplitude and wavelength, and the original lamination is almost obscured (Figs. 5D, F). A single deformed interval generally shows different fold geometries, frequencies and intensity of deformation across a full spectrum, from gentle folding to convolution (Figs. 5D, E). Folds are commonly associated with brecciated oil shales (Fig. 4), load structures (Figs. 5D, E, 6A–D) and microfaults (Figs. 5D, 6A).

3.1.1.2. Deformation mechanism

Folds point to compression as the dominant type of stress at the scale of the deformation and plastic behaviour of the sediment while it had a quasi-solid state and deformed with or without loss of shear strength through liquefaction (e.g., Allen, 1982; Mills, 1983; Owen, 1987). The lack of truncation by erosive lake floor processes and the common gradual transition into undeformed beds both upwards and downwards point to intrastratal deformation. Soft-sediment deformed oil shales in the Green River Formation have been attributed to mass wasting processes, such as slumps (e.g., Tänavsuu-Milkeviciene and Sarg, 2012), and in-situ processes (Törő and Pratt, 2015a, b). Due to the limited observations, both scenarios might be plausible, but the absence of evidence for a single transport direction and the inferred intrastratal origin favours the latter. Deformed organic-poor silty mudstone and tuff interbeds, hosted by undeformed laminated oil shale, indicate their higher liquefaction potential. Similarly, muddy interbeds in the Wilkins Peak Member generally show chaotic deformation and disruption pointing to lower resistance to liquefaction (e.g., Berra and Felletti, 2011) or thixotropic behaviour (Picard and High, 1972; Chen and Lee, 2013; Törő and Pratt, 2015a). Coarser grained, silt- and sand-rich intervals underlying plastically deformed beds, in turn, typically show the predominance of brittle structures, suggesting a more consolidated or confined state. Heterogeneity in the mechanical properties of the

sediments was originated from variable organic content, grain size, cementation, mineralogical composition, permeability (related to different grain size), and water content (e.g., Bouchette et al., 2001; Alfaro et al., 2002; Jewell and Ettensohn, 2004; Spalluto et al., 2007; El Taki and Pratt, 2012; Törő and Pratt, 2015a, b). Early cementation of these carbonate-rich sediments at shallow burial depth increased lithologic complexity and sediment cohesiveness by partial and heterogeneous cementation (e.g., Pratt, 1998b, 1999; Bouchette et al., 2001; Chen et al., 2009; Chen and Lee, 2013) and played an important role in the development of the development of sealing lithologies close to the sediment-water interface supporting the liquidization of buried sediments. Irregular fold morphology, thus, was likely the result of this heterogeneity (e.g., Rodríguez-Pascua et al., 2000), changes in rheology during the deformation event, the chaotic nature of stresses responsible for their generation, and superimposed deformation events (e.g., Montenat et al., 2007; El Taki and Pratt, 2012).

3.1.2. Load structures

3.1.2.1 Description

Mudstone interbeds within the folded intervals of the Wilkins Peak Member are often structureless, irregular in thickness, and foundered into the underlying laminae (Figs. 5E, 6A–D). Load structures are up to 5 cm in diameter, filled by a mixture of fragmented material with a muddy matrix, and spaced at irregular intervals 1–10 cm apart (Figs. 6A–E). They have a sharp but irregular lower boundary, with small-scale undulations and irregularities superimposed on the larger-scale structures, and the underlying laminae usually exhibit folding and thinning or truncation. Microfaults and small-scale sedimentary dikes are commonly associated with load structures (Figs. 6A–C, E). Load structures 2–5 cm in diameter are also present at SC within laminated fine-grained sandstone (Fig. 6F). Internally the lamination is parallel and follows around the base of the structure.

3.1.2.2. Deformation mechanism

Load structures are typically thought to form passively as a response to a density contrasts between two unstable sediment layers, or uneven loading after liquefaction (Owen, 1996, 2003). However, unlike in many previously reported examples (e.g., Anketell et al., 1970; Allen, 1982; Rossetti, 1999; Rodríguez-Pascua et al., 2000; Alfaro et al., 2002; Owen, 2003; Neuwerth et al., 2006; Berra and Felletti, 2011), a reverse density gradient is lacking between these structures and the host sediment. Load structures associated with ductile deformation unrelated to gravity likely formed by the downward squeezing of the disrupted, partially liquefied or thixotropic muddy interbeds into underlying silty sediments. At SC, load structures in sandstone with uniform grain size indicate that deformation was governed by rheological differences due to heterogeneous cementation (Owen, 2003), or to a gradient in bulk density that exists during the liquefaction state and is not detectable at the end of the process (e.g., Allen, 1982; Moretti et al., 1999).

3.1.3. Hybrid brittle–ductile structures

3.1.3.1. Description

Most deformed horizons in the Wilkins Peak Member show a combination of soft-sediment deformation together with brittle failure, that is, folds, microfaults, fault-related folds, load structures, sedimentary dikes and breccias in intervals up to 80 cm thick (Figs. 7A–H). These hybrid structures show great variability in a single layer and pass laterally into one another, confirming a common origin. A full spectrum in intensity of deformation is present, from simple, small-scale folds to contorted beds including high-amplitude folds and breccias (Figs. 7D–G). There is also an upward vertical transition from undisturbed laminated sediments into folded and faulted layers, overlain by brecciated sediment (Fig. 7C). The latter resemble ‘mixed-layers’ described by Marco and Agnon (1995) and Rodríguez-Pascua et al. (2000). In all cases, the base and top of the deformed horizons are gradual or sharp but not erosive, and they are overlain by undeformed beds of the same lithology.

Comparable structures, with similar crumpling, irregular folding and brittle deformation, have been described previously in a variety of fine-grained facies (Van Houten, 1962; Rodríguez-Pascua et al., 2000; Bachmann and Aref, 2005; Arnaud and Eyles, 2006; Montenat et al., 2007; Moretti and Sabato, 2007; Berra and Felletti, 2011; Martín-Chivelet et al., 2011; Arnaud, 2012; El Taki and Pratt, 2012; Chen and Lee, 2013).

3.1.3.2. Deformation mechanism

The association of folds, microfaults and sedimentary dikes in hybrid structures implies intrastratal liquefaction and synchronous fluidization in sediments with a transitional state between brittle and ductile behaviour. Similar to irregular folding, anisotropic rheological attributes of the host sediment favoured the formation of hybrid brittle–ductile features and incomplete liquidization, and more chaotic deformation occurred in beds with higher liquefaction potential.

3.1.4. Pinch-and-swell structures

3.1.4.1. Description

Pinch-and-swell structures (or ‘loop bedding’; Calvo et al., 1998) are present only in oil shale deposits of the Tipton Member (Figs. 8A, B). Here, irregular thickening and thinning of the lamination results in a morphology which resembles a series of loops or links of a chain, with or without disruption. Individual features are usually only a few millimetres thick and rarely larger than a centimetre. Deformed intervals, however, may reach up to several decimetres thick and listric microfaults commonly accompany them. Pinch-and-swell structures are common in all subbasins of the Green River Formation (Bradley, 1930, 1931; Cole and Picard, 1975; Dyni, 1981; Grabowski and Pevear, 1985; Törő and Pratt, 2015b).

3.1.4.2. Deformation mechanism

Pinch-and-swell structures in oil shales, akin to boudinage features, point to intrastratal ductile deformation in semi-cohesive sediment primarily due to tensile stresses arising from bedding-parallel shearing at the lamina scale (Calvo et al., 1998; Rodríguez-Pascua et al., 2000). Associated microfaults indicate concomitant brittle behaviour, and deformation at the brittle–ductile field boundary. Morphological variability, and the presence or absence of microfaults, shows that deformation was influenced by the contrasting composition and rheology of the deformed laminae.

3.2. Brittle Deformation Structures

3.2.1. Microfaults

3.2.1.1. Description

Microfaults occur in various lithologies, including oil shales of the Tipton Member (Fig. 9A), and parallel-laminated mudstones and siltstones and cross-laminated sandstones of the Wilkins Peak Member (Figs. 9B–E). In the latter, they typically cut the coarser beds and are overlain by structureless mudstone (Figs. 9B, D, E). Faulted beds contain variably dipping high-angle normal faults, reverse faults and thrust faults, and all of these may occur within a single deformed interval. Generally faults have an irregular spacing, from few millimetres up to 25 cm apart, and bedding is offset by ~0.1–40 cm (Figs. 9A, D, E). In some cases, subvertical normal faults are clustered and display domino-style geometry (Fig. 9B). Others are characterized by staircase geometries, with ragged or segmented fault plane surfaces and changing dip angles (Figs. 9C, E). Faults may be the only deformation features present in a disturbed interval (Fig. 9A), where they die out both upwards and downwards into undeformed strata. Others form part of hybrid brittle–ductile structures superimposed on folds or load structures (Fig. 9B), or occur with sedimentary dikes (Figs. 9C, D). Similar microfaulted beds have been widely documented (e.g.,

Pratt, 1994; Kahle, 2002; Neuwerth et al., 2006; Fortuin and Dabrio, 2008; Berra and Felletti, 2011; El Taki and Pratt, 2012; Törő and Pratt, 2015a).

3.2.1.2. Deformation mechanism

Microfaults indicate brittle failure of stiff, perhaps even indurated sediments as a result of elevated stresses (Owen, 1987). Variable fault types and geometries within a single interval show that fault geometry was driven by the rheological anisotropy under a combination of tensile, compressive and shear stresses (Price, 1966; Davison, 1987; Peacock and Sanderson, 1992; Pratt, 1998b; Kahle, 2002; Ferrill and Morris, 2003). Faults die out both upwards and downwards indicating intrastratal deformation, where faulting affected only certain intervals with a susceptible rheological state (Pratt, 1994; Kahle, 2002; El Taki and Pratt, 2012; Törő and Pratt, 2015a). Microfaults, even those associated with folds, indicate rheological partitioning, where plastic and brittle deformation happened simultaneously. This could happen over the course of the deforming event if sediments became more cohesive which, in rheological terms, is called 'strain hardening'. Alternatively, in some cases, faulting may have occurred during a later event affecting the previously deformed features.

3.2.2. Breccias

3.2.2.1. Description

Brecciated intervals in the Wilkins Peak Member are either common throughout the studied section (SC), or occur only in the upper part of the first cycle (e.g., SJC, MFC, FC) (Figs. 3, 10A–F). These are generally mound- or irregular-shaped bodies up to 1 m in diameter (Figs. 10C–F) and rarely stratiform (Figs. 10A, B), and all are traceable laterally for several tens of metres. Breccia lenses may have a sharp boundary with undeformed beds (Figs. 10B), but gradual transitions are also common and

marked by progressive brecciation (Figs. 10C–F). Despite the restricted size of individual bodies, the same interval typically hosts several breccia lenses along with other deformation features.

Brecciation typically affected silt- and sandstone-dominated lithologies, variously interbedded with mudstone. Thus, they consist of platy, angular to subangular clasts of these lithologies that are up to 10 cm in size, and individual clasts can be folded and faulted. The clasts show no preferred orientation, and flat-lying, inclined, or edgewise attitudes coexist. Occasionally, clasts follow curvilinear features and breccias are similar to sedimentary dikes (Figs. 10C, D). The matrix is usually a structureless, mud–silt mixture.

At SC several mudstone layers interbedded within fine-grained sandstones show fragmentation, folding, and various amounts of transportation (Fig. 10B). The sandstone beds are structureless, and sand fills the voids between the mudstone fragments. In places the sand continues downwards or upwards forming small-scale sedimentary dikes resembling Type 4 dikes (Fig. 10B). On bedding plane surfaces they show an irregular polygonal pattern and the sand-filled dikes superficially resemble desiccation cracks. The fragments are variable in size, but generally 1–6 cm in diameter. Some rounded or slightly folded fragments are dislodged and incorporated within the underlying sandstone where they float in a structureless sandy matrix. Load structures are commonly found within these intervals.

3.2.2.2. Deformation mechanism

Breccias indicate disruption of cohesive sediments without significant transportation as shown by the angularity of the clasts. Folding of the host strata and load structures indicate the plastic behaviour of the semi-cohesive sediments during brecciation. Based on the irregular geometry of the breccia bodies, their sharp and non-erosive tops, and in some cases the transitional boundaries to undeformed host sediments, brecciation occurred in situ under shallow burial depth. Their mud-rich matrix implies that brecciation was the result of selective fluidization of the less cemented muddy intervals and the synchronous disruption of the more indurated silty and sandy units due to elevated

pore pressures (e.g., Van Houten, 1964; Kahle, 2002; Montenat et al., 2007; Berra and Felletti, 2011; El Taki and Pratt, 2012; Chen and Lee, 2013). Occasionally, brecciation was followed by sediment remobilization and injection which formed dike-like features. In these cases, a sequence of closely spaced events can be inferred: (1) folding of the semi-lithified bed, with or without liquefaction; (2) disruption, brecciation, and homogenization; and (3) remobilization or injection, plus shearing. This, together with transitional boundaries, implies an intrastratal origin. Assuming similar rheological properties, complete disruption of the semi-lithified sediment layers and subsequent remobilization likely point to higher deformation intensity compared to other deformation features, such as folds, load structures, or microfaults (e.g., Rodríguez-Pascua et al., 2000; Berra and Felletti, 2011). Breccia layers with massive sandy matrix at SC are interpreted as autoclastic formed by the in-situ brecciation of an indurated mudstone or silty mudstone layer (e.g., Montenat et al., 2007). Here, liquidization and homogenization resulted in the structureless nature of the sandstone.

3.2.3. Oil shale breccias

3.2.3.1. Description

Brecciated oil shale intervals are hosted by undeformed oil shale and occur in all cores (Figs. 4, 11A–F), and have been also noted in other subbasins of the Green River Formation (e.g., Bradley, 1931; Dyni and Hawkins, 1981; Tänavsuu-Milkeviciene and Sarg, 2012). The breccias are characterized by a structureless organic-rich mudstone matrix and folded laminated fragments with variable organic content, up to 3 cm in length, that are aligned parallel to the bedding plane with no grading (Figs. 11A, D, F). These intervals are up to 75 cm thick and they usually show sharp and planar upper and lower boundaries. Gradual disruption at the base, top, or internally appears to occur locally (Figs. 11B, C). Deformation is more distinct where brecciation has affected laminae with variable organic content (Figs. 11C, D), whereas it is obscure in homogeneous intervals (Figs. 11A, E, F). In some cases, brecciated oil shale is associated with folded and fragmented laminae (Fig. 11E).

3.2.3.2. Deformation mechanism

Similar to breccias, oil shale breccias indicate brittle failure of indurated organic-poor laminae accompanied by the ductile, fluid-like behaviour of the organic-rich matrix. Rupture was likely associated with the thixotropic failure of organic-rich intervals, as shown by modern and ancient analogues (e.g., Picard and High, 1972; Grimm and Orange, 1997; Montenat et al., 2007). These breccias are abundant in other subbasins of the Green River Formation and are generally interpreted as mass-transport deposits (e.g., Dyni and Hawkins, 1981; Birgenheier and Vanden Berg, 2011; Tänavsuu-Milkeviciene and Sarg, 2012) or the products of in-situ disruption (Törő and Pratt, 2015a, b) similar to other published examples lacking evidence for long-distance transportation (Agnon et al., 2006; Sakaguchi et al., 2011). The restricted level of observation presented by cores prevents a definitive explanation. Nevertheless, where gradual boundaries are present, or breccias are associated with other deformation features such as folded and fragmented laminae, an in-situ, intrastratal origin is favoured. Although the lateral extent of each brecciated bed is uncertain, at least two brecciated intervals are present in each core that may be correlatable (Fig. 4). The one below the Tipton–Wilkins Peak boundary is represented by soft-sediment deformed oil shale at MFC (Fig. 3), and likely correlation of these intervals suggests a genetic relationship and a lateral extent of >38 km. If so, this points to more regional in-situ disruption, rather than localized mass-transport.

3.3. Sedimentary dikes

The term ‘sedimentary dike’ is used here to describe millimetre- to metre-scale intrusions filled with sediment, regardless of their infill or origin. Features in the smaller end of the spectrum have also been termed ‘cracks’ or ‘dikelets’. Based on their morphological characteristics, infill, and position within a single bed or a succession of beds, five main types of sedimentary dikes are distinguished here.

3.3.1. Description

Type 1 dikes are restricted to the Tipton Member where they occur as isolated, irregular-shaped blobs, 2–5 cm in diameter (Figs. 12A, B), or tabular or spindle-shaped cracks (Fig. 12C), and contain a mixture of carbonate mud and silt. The latter are up to 15 cm long and pass downwards and upwards into subtle narrow cracks that pinch out. Original lamination of the host strata is frequently folded or faulted in the vicinity of these dikes (Figs. 12A–C). Deformation structures similar to Type 1 dikes have been previously recorded in passing from oil shale deposits in several subbasins of the Green River Formation (Bradley, 1930; Picard and High, 1972; Smoot, 1983; Birgenheier and Vanden Berg, 2011; Tänavsuu-Milkeviciene and Sarg, 2012).

Type 2 dikes are linear, parallel-sided, up to 3 m long and 5 cm wide (Figs. 13A–E), and filled with homogeneous mud, silt or a mixture with sand- and pebble-sized clasts (Figs. 13F, G). Both in cross section and on the bedding plane surfaces dikes form ragged and bifurcating linear paths with sharp boundaries (Figs. 13A, B). In vertical view the dikes are straight, but at distinct lithologic boundaries they commonly show step-like lateral shifts (Figs. 13C, D). Type 2 dikes are present at two locations, MFC and RC, where they form closely spaced networks in mudstone and siltstone (Figs. 13A, B). At MFC many follow normal microfaults and penetrate silty interbeds that are locally folded (Figs. 13A, D). At RC they have two prominent orientations, $\sim 005^\circ$ and $\sim 120^\circ$, and show cross-cutting relationships with Types 3, 4 and 5 dikes. Their upper termination is generally overlain by parallel- or wavy-laminated siltstone or sandstone, or mudstone (Figs. 13A, C, E). Occasionally, the overlying bed shows disturbed lamination and brecciation. In general, dikes taper downwards, indicating that they were sourced from above. However, in many cases, the direction of tapering or a definite source interval cannot be observed.

Type 3 dikes are seldom longer than 0.5 m, and they are commonly filled with brecciated mudstone and siltstone clasts up to 4 cm across with a muddy matrix (Figs. 14A, B, D). They have an overall V-shaped cross-section, with the top up to 25 cm wide, but they frequently show ptigmatic

folding (Figs. 14A, B), or sills projecting sideways from the main dikes (Fig. 14C). The dikes are commonly rooted in an overlying source bed that shows irregular thickness and brecciation (Figs. 14A, D). Type 3 dikes frequently cross-cut silty and sandy beds that form the tops of upward-coarsening cycles, and are filled by carbonate mudstone from above, forming a mudstone 'plug' at the top of the dike, while the lower part of the dikes is filled with brecciated material (Figs. 14A, B). Multiple infill generations are also present (Fig. 14B). Many of the dikes show enhanced, silica-rich cementation that gives them a distinct orange-coloured weathering. One observed dike has laminated microbial carbonate laminae encrusting its margins followed by a brecciated infill including fragments of the microbial laminae (Figs. 15A–C). Type 3 dikes occur only at RC, and are frequently associated with Types 2, 4 and 5 dikes.

Type 4 dikes are millimetre- to centimetre-scale structures that penetrate into muddy intervals at the base or at the top of individual silty mudstone, siltstone, or sandstone beds or laminae (Figs. 16A–G). They occur as a dense network of dikes at the base and top of each lamina (Figs. 16A–E), or sparsely spaced (Figs. 16F, G). They are filled either by material of the source bed that they are attached to (Figs. 16A–E), or, less frequently, with finer grained material within a coarser grained bed, where the overlying source is missing or obscured by deformation (Figs. 16F, G). Dikes either taper downwards or upwards, but usually both directions are present (Figs. 16D, E). In cross-section they are straight and tilted (Fig. 16A) or ptygmatically folded (Figs. 11B–E), and the infill can be fragmented (Fig. 16G). Type 4 dikes are commonly associated with other deformation features, like microfaults (Figs. 9C, D) or folds and load structures (Figs. 7B, H). Where visible on bedding planes, they are classified as Type 5 dikes, indicating a common origin (see below).

Type 5 dikes are centimetre-scale, straight to curvilinear, doubly tapering or branching on bedding surfaces (Figs. 17A–I). They are seen on the top or at the base of sandstone or siltstone beds with positive relief, 1–3 mm high. There is a continuous spectrum from scarce spindle-shaped (Figs. 17C, D, I) to densely spaced triradiate forms comprising incomplete polygonal networks (Figs. 17A, B, F, G). Regardless of their spacing, dikes can have a random orientation (Figs. 17A–F, I) or can follow a

preferential direction (Figs. 17G, H), and both styles can be closely spaced stratigraphically within the same sedimentary unit (Fig. 17G). At RC they are oriented in the same direction as Type 2 dikes. In a few cases two or three generations of dikes can be distinguished based on cross-cutting relationships, and these possess more complex and irregular morphologies. Type 5 dikes have been observed only at RC and MFC, although they are also abundant locally stratigraphically higher in the Wilkins Peak Member. Occasionally, Types 4 and 5 dikes are directly related, where the latter represents the bedding plane view of the former. This relationship, however, is not always apparent in the outcrop or core.

3.3.2. Deformation mechanism

In general, the formation of sharp-walled sedimentary dikes indicates brittle failure of cohesive or semi-lithified sediments that potentially can be triggered by a number of mechanisms (e.g., Maltman, 1994; Jolly and Lonergan, 2002; Montenat et al., 2007). The structureless infill of Type 1 and Type 2 dikes indicates that they formed by the injection of fluidized sediments into fissures driven by increased interstitial pore pressure (e.g., Daley, 1971; Calvo et al., 1998; Montenat et al., 2007; Moretti and Sabato, 2007; Berra and Felletti, 2011; Hurst et al., 2011). Isolated Type 1 dikes imply dewatering, shrinkage, and segregation and remobilization carbonate mud from the surrounding sediment intrastratally (e.g., Pratt, 1998b, 1999; Törő and Pratt, 2015a). Irregular blob-like bodies filled with silty mudstone suggest that silty source beds were completely remobilized.

Based on field observations most Type 2 dikes were emplaced downwards. Where the mode of emplacement is uncertain and a source interval cannot be identified dikes might have formed by injection of material from below or sideways (e.g., Rossetti, 1999; Rodríguez-Pascua et al., 2000), or the source bed was entirely mobilized and injected. Elevated pore pressure, forceful injection, and hydrofracturing likely happened intrastratally as dikes penetrate fault planes that die out both upwards and downwards. Cross-cutting relationships with other deformation features and multiple infills indicate that the same sediments were affected by two or more deformation events at different stages of early

diagenesis. The orientation of the dikes can be related to stresses induced by the trigger mechanism or the regional stress field upon deformation (Pratt, 1998b; Jolly and Lonergan, 2002; Törő and Pratt, 2015a).

Morphological characteristics and infill of Type 3 dikes indicate that they may have formed through the passive infill of subaqueous fissures that were occasionally open on the lake floor for enough time to have microbial mats forming on their sides. Fracturing allowed infilling by concurrently brecciated material followed by background sedimentation. Alternatively, fissures formed intrastratally due to tensile stresses and elevated pore pressures at a later stage caused brecciation of their fill, fluid movement, and further sediment mobilization. Portions of the dikes may have been cavities for a period of time, allowing the growth of microbial mats. In any case, multiple fluidization events are shown by various generations of brecciated infill, which point to brittle failure due to elevated stresses after dike formation, when the infills were indurated (Pratt, 1998b, 1999). Tilting andptygmatic folding of dikes have been attributed to compaction (e.g., Hillier and Cosgrove, 2002; Parize et al., 2007). However, progressive cementation and brecciation preclude the possibility of significant compaction after burial, and the sills rooted in the folded dikes indicate that the final morphology is the result of the dike-forming process (Törő and Pratt, 2015a).

Types 4 and 5 dikes are seen to frequently occur together, depending on the nature of the outcrop, and, thus, they are considered genetically related. Various mechanisms suggested to explain them, mainly in siliciclastic settings, including subaerial desiccation (with or without the role of evaporites) (e.g., Kidder, 1990; Astin and Rogers, 1991), subaqueous shrinkage of muddy sediments due to salinity changes (syneresis) at the sediment–water interface or intrastratally (e.g., Van Houten, 1964; Donovan and Foster, 1972; Plummer and Gostin, 1981), shrinkage of thin microbial mats subaerially, subaqueously or intrastratally (e.g., Parizot et al., 2005; Eriksson et al., 2007a, b; Harazim et al., 2013), or post-depositional intrastratal dewatering of argillaceous sediment caused by compaction (Kidder, 1990; Tanner, 1998) or seismic shaking (Pratt, 1998a). In the Green River Formation similar features have been

reported from all the subbasins, and have been interpreted as linear shrinkage cracks formed by desiccation (Bradley, 1930; Smoot, 1983), gravitational forces (Picard, 1966), dewatering (Birgenheier and Vanden Berg, 2011; Tänavsuu-Milkeviciene and Sarg, 2012), or shrinkage combined with sediment injection (Törő and Pratt, 2015a, b). First, it should be noted that salinity fluctuations related to freshwater input by groundwaters or rivers are unlikely in a hypersaline, stratified portion of the lake far from the shoreline, and secondly the original mineralogical composition of the deformed beds was not particularly argillaceous such that passive chemical shrinkage is unlikely (cf. Bradley, 1931; Burst, 1965; Roehler, 1992a; Demicco and Hardie, 1994). Genesis related to the presence of microbial mats is excluded as no other structures indicating their presence have been observed (cf. Noffke, 2009). Moreover, some of the dikes taper upwards, indicating upward intrastratal injection into overlying beds under elevated pressure (e.g., Scott and Price, 1988; Obermeier, 1996; Rodríguez-Pascua et al., 2000; Hurst et al., 2011). Downward-tapering dikes clearly penetrated into fissures formed by dewatering and shrinkage of the muddy interbeds (Pratt, 1998a). Types 4 and 5 dikes are interpreted to represent a continuum in response to deformation processes, where vertical Type 4 dikes formed: (1) in the presence of thicker laminae which allowed remobilization of adequate sediment; (2) under greater pore pressures which was accompanied by shrinkage; or (3) in less lithified sediments allowing injection and shrinkage. In other cases, sediment remobilization was dominantly along bedding surfaces, which resulted in the formation of spindle shaped Type 5 dikes.

In previous reports on the Green River Formation the term 'mudcracks' has been extensively applied to a variety of sedimentary features including isolated (Type 1), linear (Type 2), V-shaped or folded (Types 3 and 4), or spindle-shaped (Types 4 and 5) dikes in lithologies ranging from littoral–sublittoral, siliciclastic or calcareous rippled sandstones, siltstones, silty carbonate mudstones and limestones, to profundal laminated carbonate mudstones and oil shales (e.g., Bradley, 1930, 1964; Eugster and Hardie, 1975; Smoot, 1983; Roehler, 1991b, 1992a). Typically, these features have been taken as evidence of subaerial exposure, even though in most cases their morphology on bedding plane surfaces and the characteristics of their infill are clearly dissimilar to true desiccation cracks (Allen, 1987;

Törő and Pratt, 2015a, b; Törő et al., 2015). Moreover, the host strata lack accompanying sedimentary signatures consistent with subaerial exposure or pedogenesis, such as caliche nodules, mottled and variegated mudstones, or oxide haloes and crusts. Thus, a desiccation origin is rejected based on the inferred deeper water depositional and paleogeographic setting, and the vertical extent, uniform width, and linear nature of the dikes, plus the absence of horizontal lamination or grading in the infilling material as would be expected if the crack were open and filled by wind or fluvial processes (cf. Allen, 1987).

3.4. Deformed tuffs

3.4.1. Description

Tuff beds frequently have sharp but irregular, wavy upper and lower boundaries and those within the Tipton Member are typically massive with oil shale clasts locally, lack any grading and show irregular folding, fluid-escape structures and brecciation (Figs. 18A–C). Centimetre-scale sedimentary dikes, resembling Type 4 dikes, frequently originate from the base of the tuff beds and penetrate downwards into the underlying oil shale whose lamination is folded around them (Figs. 18D, E). These dikes are usually tilted and variably folded, and similar to those reported from the Uinta Basin (Törő and Pratt, 2015b). Some tuff-filled dikes, however, have no clear connection to their source bed and appear isolated (Fig. 18E).

3.4.2. Deformation mechanism

Deformed tuff beds have been widely documented, especially in the Uinta Basin, but only a few have noted the presence of associated dikes (Cashion, 1967; Griggs, 1968; Surdam and Stanley, 1979; Grabowski and Pevear, 1985; Törő and Pratt, 2015b). Disturbed bedding was related to surface or subsurface remobilization and “plastic flowage” (Bradley, 1931; Cashion, 1967; Remy, 1992). The dikes,

in turn, have been attributed to passive subaerial infill of desiccation cracks (Surdam and Stanley, 1979), or remobilization and injection of colloidal-like tuff upwards during burial (Cashion, 1967; Griggs, 1968). The structureless nature of the tuffs and the association of irregular folds and fluid-escape cusps point to liquefaction with the development of fluidization due to elevated pore pressures (Owen, 1987, 2003). Irregular upper and lower boundaries and the lack of erosion imply that the deformation was intrastratal. The relatively undeformed nature of the host oil shale demonstrates that the tuff beds were more prone to liquefaction, and this is due to their granular nature. Downward-tapering of tuff-filled dikes, similarly to Type 4 dikes, suggest that liquefaction was combined with dewatering and shrinkage of the underlying oil shale (e.g., Törő and Pratt, 2015a, b).

3.5. Deformed trona

3.5.1. Description

White, fibrous trona, referred to as ‘satin spar trona’ (e.g., Jagniecki and Lowenstein, 2015), occurs at the base of the first trona bed in the BF1 core where it fills a subhorizontal 1–1.5 cm wide fracture (Fig. 19A) as it does elsewhere within the Wilkins Peak Member (Birnbaum and Redlick, 1987; Leigh, 1991; Wiig et al., 1995). Trona intervals above show brecciation, with coarse crystalline, fibrous, and bladed trona fragments reaching 4 cm in diameter, and incorporating angular, irregular-shaped, or folded mudstone clasts (Figs. 19A, B). Mudstone interbeds also display ductile and hybrid brittle–ductile deformation, with irregular folds and microfaults (Fig. 19B). Trona deposits of the Wilkins Peak Member typically exhibit fine-grained and acicular crystals, or radiating and bladed aggregates (e.g., Deardorff and Mannion, 1971; Roehler, 1993; Pietras and Carroll, 2006; Jagniecki and Lowenstein, 2015), and the occurrence of brecciation has been reported only in passing.

3.5.2. Deformation mechanism

Fracture-filling trona has been interpreted as the result of hydrofracturing and trona precipitation from saturated brines due to post-depositional overpressure (Birnbaum and Redlick, 1987; Leigh, 1991). Remobilization, disruption, and brecciation of trona beds with slump structures, evaporite-filled sedimentary dikes, and the disruption of mudstone interbeds, have been mentioned only in passing in previous studies (Deardorff and Mannion, 1971; Birnbaum and Redlick, 1987; Leigh, 1991; Wiig et al., 1995; Dyni, 1996; Boni and Atkinson, 1998). Brecciation of the trona layers indicates brittle failure of the trona with synchronous brittle–ductile deformation of the semi-cohesive mudstone interbeds. Irregular and folded mudstone fragments originate from the disruption of the host mudstone and indicate that some layers were stiff and some were still ductile at the time of deformation. Whether or not the trona was deformed by mass-transport or by in-situ processes cannot be determined on the basis of core-scale observations (see Trigger Mechanism). However, breccia composition and the lack of littoral lithologies point to a nearby source.

4. Discussion

4.1. Trigger mechanism

Majority of the deformation features are attributed to elevated pore pressure and external compressive and tensile stresses in unconsolidated to semi-consolidated deposits intrastratally with or without liquefaction and fluidization (Fig. 20). The overall morphology was controlled by the horizontal or vertical shear stresses, gravitational body force, or unstable density gradient, and sediment behaviour was controlled by the rheological properties during deformation as well as the intensity of the stresses imposed. There are a number of possible trigger mechanisms that could conceivably generate synsedimentary deformation structures, such as strong wave and tidal action, rapid sediment loading, drag by subaqueous currents, ice-induced stresses, earthquakes, slumping, and groundwater or gas seepage (e.g., Owen, 1987, 2003; Moretti et al., 1999; Owen and Moretti, 2011). Similar features can

ostensibly be created by different triggers, which has led to uncertainty in interpretation (e.g., Jones and Omoto, 2000; Alfaro et al., 2002; Montenat et al., 2007; Owen et al., 2011).

In terms of endogenic triggers, i.e. intrinsic to the sedimentary environment, the mid-latitude paleogeographic position and the warm climate of the early to middle Eocene (e.g., Roehler, 1993; Zachos et al., 2001) exclude the possibility of ice-induced deformation. The interpreted depositional setting and the lack of turbidites, slumps, or channel deposits also indicate that rapid depositional events, overloading, or drag by subaqueous currents can be excluded. Deformation related to concentrated vertical fluid movement (water or gas), including artesian springs (e.g., Li et al., 1996; Massari et al., 2001; Ledésert et al., 2003), is also unlikely, as the volcanic centres that could have sourced escaping gases were distant (e.g., Roehler, 1993; Carroll et al., 2008), and the breccias and other deformation features do not indicate vertical paths for fluid movement. Shear stresses associated with tidal waves can cause ductile deformation (e.g., Wells et al., 1980; Greb and Archer, 2007), but the morphology of these deformation structures contrasts with those described here, and significant tidal action is unlikely in lakes (e.g., Talbot and Allen, 1996; Renaut and Gierlowski-Kordesch, 2010). The depositional setting of the outcrops and the general rarity of evaporites preclude the possibility of disruption and deformation of laminae caused by growth of evaporite minerals, even if now vanished. Small evaporite (trona?) casts in the Tipton Member are not associated with deformation features. Moreover, most of these triggers would generate only local deformation features and cannot explain the laterally extensive and often vertically extensive nature of the structures.

Liquefaction of sediments can be related to the action of breaking or pounding waves (Seed and Lee, 1966; Sangrey et al., 1969; Seed and Rahman, 1977; Dalrymple, 1979) and there are several cases where wave-related stresses have been invoked for sedimentary deformation (e.g., Kerr and Eyles, 1991; Martel and Gibling, 1993; Molina et al., 1998; Bouchette et al., 2001; Alfaro et al., 2002; Chen et al., 2009). The absolute depth of Lake Gosiute is not known, but it probably reached tens or perhaps more than a hundred metres in the centre (e.g., Bradley, 1926, 1929; Surdam and Stanley, 1979;

Steidtmann et al., 1983; Roehler, 1990), with a very gently dipping slope (0.5–1° or less; e.g., Pipiringos, 1955; Bradley, 1964; Roehler 1990). Although the lake had a significant fetch, conditions were most likely insufficient to produce large waves capable of inducing liquefaction at depth (cf. Nataraja and Gill, 1983). Furthermore, the liquefaction features are not associated with storm beds (cf. Martel and Gibling, 1993; Alfaro et al., 2002), and the dominant parallel lamination of the host facies does not indicate significant wave energy. Mass-transport deposits related to endogenic triggers, such as oversteepening or overloading, are excluded based on fine-grained lithology and the insignificant dip of slope (e.g., García-Tortosa et al. 2011; Törő and Pratt, 2015a). Additionally, deformation features show no uniform vergence or unidirectional axial planes of folds that would point to downslope movement (e.g., Strachan, 2002; Alsop and Marco, 2012).

Because all the endogenic processes can be rejected the sedimentary deformation structures must be related to an external, or exogenic, trigger. They are best explained as the result of shear stresses and pore pressure increase in-situ induced by seismic shaking generated by syndepositional earthquakes. Thus, these sedimentary deformation features are interpreted as seismites (Seilacher, 1969, 1984). The chaotic compressive and tensile stresses related to the repeated passage of seismic waves combined with the rheological heterogeneity explains the unoriented and contorted nature of the folded, convoluted structures, and microfaults, the recurrence of laterally extensive (>50 km) deformed intervals that are stratally bound by undeformed beds, and the combination of brittle and ductile deformation. Furthermore, syndepositional tectonic activity in the basin has been inferred (Steidtmann et al., 1983; Roehler, 1992b; Pietras et al., 2003; Pietras and Carroll, 2006). Turbidites are widely used as paleoseismic indicators in lacustrine and marine settings (e.g., Adams, 1990; Becker et al., 2005; Patton et al., 2015; Van Daele et al., 2015). Although thin normal graded silty mudstone beds interbedded within oil shale might be interpreted as distal turbidites (Walker, 2008), their near absence in the cores and outcrops in the study area shows that most clastic sediment remained near the lake margin of the carbonate-dominated Lake Gosiute, and the extremely gentle paleoslope was not conducive to sediment transport towards the depocentre.

Deformation of evaporites, in particular, has been attributed to a variety of processes, such as volume expansion due to crystal growth especially in sulfate evaporites (e.g., Collins, 1984; Schreiber and Helman, 2005), or collapse related to evaporite dissolution (e.g., Friedman, 1997; Eliassen and Talbot, 2005; Broughton, 2013), as well as earthquakes (Bachmann and Aref, 2005; El Taki and Pratt, 2012). Brecciation of trona beds in the Wilkins Peak Member has been attributed to recrystallization (Deardorff and Mannion, 1971), reworking by floods (e.g., Birnbaum and Redlick, 1987; Leigh, 1991), and disruption by upward-directed mud injection caused by overpressure (Boni and Atkinson, 1998). The subaqueous depositional setting of the trona beds excludes the possibility of subaerial reworking, and sediment composition militates against the possibility of reworking by gravity flows originated from the littoral zone. Volume changes related to crystal growth, recrystallization, or replacement of earlier precipitated evaporites is minimal (e.g., Birnbaum and Redlick, 1987; Pietras and Carroll, 2006; Jagniecki and Lowenstein, 2015) and cannot account for the scale of deformation described here. Collapse related to evaporite dissolution results in cavity formation and extensive gravity-controlled brecciation of the overlying beds. However, structures typical of collapse, such as geopetal indicators, inverse grading, and dissolution fronts, are lacking, the overlying and interbedded units are undeformed or show mixed brittle–ductile deformation, and none of the observed features indicate downward movement. Additionally, dissolution features in thick trona beds of the Wilkins Peak Member show thinning coupled with downwarped shale and trona beds, rather than brecciation (e.g., Deardorff and Mannion, 1971). Deformed trona described here are more akin to irregular trona breccia beds illustrated by Leigh (1991), and are interpreted as the result of in-situ disruption with or without short-distance translation. Alternatively, they might represent disruption caused the forceful upward injection of fluidized mud (“mud squirts” of Boni and Atkinson, 1998). Although this was not observed in the cores, fluidization of the underlying carbonate mud as well as in-situ disruption were likely the result of seismic disturbance. Synchronous deformation features present in other cores and outcrops also suggest a regional, earthquake trigger. Similarly, hydrofracturing and the formation of fracture-filling fibrous trona are attributed here to earthquake-triggered overpressure and precipitation after burial.

4.2. Earthquake characterization

4.2.1. Earthquake magnitude and source

The relationship between earthquake parameters, like magnitude and epicentral distance and the stratigraphic occurrence of fluidization features like sand blows have been employed to determine the magnitudes and intensities of Holocene earthquakes (e.g., Allen, 1986; Leeder, 1987; Audemard and De Santis, 1991; Obermeier et al., 1993; Obermeier, 1996; Scott and Price, 1988; Galli, 2000; Obermeier et al., 2001; Castilla and Audemard, 2007). These examples suggest that magnitude 5–5.5 and intensity VII MCS represents a minimum threshold needed for deformation to occur in near-surface subaqueous sediments. Additionally, an earthquake with $M=7$ is capable of producing liquefaction as far as 70–100 km from the epicentre. The array of ancient sedimentary deformation features can only be used locally as indication of earthquake magnitude because initial properties of the sediment are more important than the actual magnitude (e.g., Rodríguez-Pascua et al., 2000; Alfaro et al., 2010; Berra and Felletti, 2011). Additionally, most of the seismite types that formed in sediments under a water column have yet to be documented in Recent examples. Also, the minimum threshold could be higher because of the increasing tensile strength of carbonate sediment undergoing cementation.

The closely spaced nature of seismite-bearing horizons at the Tipton–Wilkins Peak transition indicates frequent seismic events from a relatively nearby source. In the study area there are two fault zones 100–150 km away that were certainly active during deposition: the Wind River Thrust system (105–160 km to the north; e.g., Steidtmann et al., 1983; Pietras et al., 2003) and the Uinta Thrust system (15–65 km to the south; e.g., Roehler 1992b, 1993; Bradley, 1995). Additionally, syndepositional faulting and growth of the Rock Springs Uplift has also been suggested (Bradley, 1964; Bradley and Eugster, 1969; Goodwin, 1971; Smith et al., 2014a). These structures could have generated earthquakes of $M \geq 5$ based on recent observations in compressional tectonic settings (e.g., Castilla and Audemard, 2007; McCalpin, 2009). The geographic distribution of the seismites, which can point to the source (e.g.,

Leeder, 1987; Ettensohn et al., 2002), is in the Bridger Basin more or less uniform such that specific tectonic elements are not implicated (Fig. 21). This might be due to the close proximity of several fault systems that may have been active, and site effects influencing seismite formation.

Our study sites were far apart but their distribution was not wide enough to indicate a lateral zonation of the deformation style (cf. Ringrose, 1989). However, deformation features of the northern locations (all outcrops except RC) show an increase in the relative proportion of breccias towards the south (SC) (Fig. 3). This change coincides with an overall coarsening of the sediments and an increase in cycle thickness. This might have influenced early diagenetic processes and the rheological properties of the sediments, and increase in stiffness could explain this trend in sediment behaviour. Alternatively, breccias may point to higher deformation intensity (e.g., Berra and Felletti, 2011), and increase in their abundance can be related instead to the change of earthquake parameters, such as epicentral distance or magnitude. The latter is supported by the dominance of dikes at RC, which is more likely related to increased deformation intensity as the northern locations have a similar lithology (e.g., Rodríguez-Pascua et al., 2000; Hilbert-Wolf et al., 2009; Berra and Felletti, 2011). Consequently, the most probable source for the earthquakes was faulting along the Uinta Thrust System to the south, but movements on buried faults related to the Rock Springs Uplift cannot be discounted.

To estimate minimum magnitudes and intensity for the paleoearthquakes, we used equations in Ambraseys (1988), Galli (2000), and Castilla and Audemard (2007) which are based on empirical relationships between magnitude or intensity, and maximum distance of deformation feature from the epicentre or fault. Using distance of the Uinta Thrust System and the WM1 core, ~65 km, these gave minimum magnitudes between 6.4 and 7.1, and minimum epicentral intensity (Mercalli-Cancani-Seiberg) of ~9.4, which are used to place crude limits on possible earthquake parameters.

4.2.2. Temporal distribution and recurrence periods

The deformation structures are clustered in five to seven intervals in outcrops of the northern part of the study area, while their number increases to fourteen at RC and up to twenty-two in the cores (Figs. 3, 4). Observations made in cores could lead to underestimation of seismic events if deformed horizons are laterally impersistent and not intersected (e.g., Ezquerro et al., 2015). The differences between localities must be related in general to changes in factors controlling seismite formation, including earthquake magnitude, intensity and duration of ground motion, and confining pressure related to water depth and burial depth at site, along with the physical attributes of the sediment.

In the simplest sense, each deformed interval records a single earthquake, although it is clear that some, such as hybrid structures, breccias and Type 3 dikes, do record a second or even a third event superimposed while the sediment was still susceptible to deformation. On the other hand, it is possible a single earthquake could deform two or more closely spaced intervals separated by intact beds, if rheology was stratigraphically heterogeneous. A simple tally of deformed horizons in both outcrop sections and cores indicates a minimum of 15 to 22 distinct events over a stratigraphic interval of 15–26 m.

Determining the recurrence pattern of earthquakes, especially in Quaternary deposits (e.g., Sims, 1975; El-Isa and Mustafa, 1986; Marco et al., 1996; Migowski et al., 2004) has relied on features that formed at the sediment–water interface, which makes the dating of individual seismic events more accurate. In the studied interval, most seismites formed intrastratally, at some point after burial, which increases the degree of uncertainty in the timing.

The timing of strong earthquakes is estimated using average depositional rates from previous reports (e.g., Smith et al., 2008, 2014a) and depositional rates calculated from two dated tuffs coupled with published stratigraphic data (Pietras, 2003; Walker, 2008) (Table 1). The Sheggs tuff (52.22 ± 0.09 Ma; Smith et al., 2008, 2010) is located close to the boundary of the Sheggs and Rife beds of the Tipton Member, and the Firehole tuff (51.41 ± 0.21 Ma) occurs in the lower Wilkins Peak Member (Fig. 2A). Due to the absence of major unconformities and paleosols in the studied interval, continuous sediment

accumulation was assumed, although rate varied spatially. Calculated depositional rates are in good agreement with the 110 mm/k.y. given by Smith et al. (2008), so depending on thickness the studied outcrops and cores represent between ~45 and 236 k.y. of lacustrine sedimentation. Outlying values at RC may be due to higher actual sedimentation rates or uncertainty in correlation.

Apparent recurrence periods (Ezquerro et al., 2015) were calculated in two ways: (1) using the total number of observed seismites (Table 2); and (2) by the transformation of stratigraphic thicknesses between two seismite horizons into temporal intervals (cf. Marco et al., 1996) (Table 3). The latter was performed only on cores, as they provide the most complete record of seismic events.

The calculated apparent recurrence periods are variable and depend on the number of observed deformed horizons and the inferred depositional rates (Tables 2, 3). Moreover, the thickness of strata between seismites is not constant, resulting in variable intervals times and large standard deviations. Nevertheless, the calculated mean recurrence period is ~8.1 k.y. and cores give a broadly uniform apparent recurrence period of 5.8–8.1 k.y. depending on the location.

Clustering of events is noticeable in the three cores both in the stratigraphic distribution of seismites (Figs. 4, 22) and the distribution of temporal intervals (Figs. 23A, B). Additionally, each core suggests a similar clustering, where two active periods with relatively short recurrence times (~2–4 k.y.) are followed by relative tectonic quiescence and long interval times (~10–25 k.y.). This pattern is independent of the applied depositional rates. Closely spaced deformed beds shown by very short temporal intervals (<1 k.y.) indicate two closely spaced events or deformed intervals generated by the same event.

Lake Titicaca in the Bolivian Andes has been considered a modern analogue for this part of the Green River system both geomorphologically and tectonically (Smith et al., 2014a). In the 50–100 km vicinity of the lake the recurrence time of earthquakes with $M \geq 5$ averages one year based on the USGS earthquake database (<http://earthquake.usgs.gov/earthquakes/>). Thus, it is reasonable to expect that broadly comparable large-magnitude earthquakes could have affected the Green River lake system

during tectonically active episodes. This also shows that numerous earthquakes with magnitudes insufficient to induce deformation might have occurred, but were not recorded.

4.3. Implications for lake evolution

The Green River Formation documents an intricate history of lacustrine deposition in that was the result of the complex interplay between lake sedimentation and climate, tectonics, and geomorphologic changes (Bradley, 1929, 1964; Surdam and Stanley, 1980; Roehler, 1993; Rhodes et al., 2002; Carroll et al., 2006, 2008; Smith et al., 2008, 2014a, b; Doebbert et al., 2010, 2014). Changes in the paleohydrological conditions of Eocene Lake Gosiute are marked by regional lithologic boundaries that divide distinct lake-stages and can be traced across the basin (e.g., Roehler, 1992a, b; Carroll and Bohacs, 1999; Bohacs et al., 2000; Carroll et al., 2008; Doebbert et al., 2010). Short-term climatic changes are thought to be responsible for the cyclic nature of the deposits (Fischer and Roberts, 1991; Roehler, 1993; Aswasereelert et al., 2013), while syndepositional tectonic activity along basin-bounding structures had longer term effects, as it caused crustal flexure which continuously shifted the successive depocentre and the regional distribution of depositional environments (e.g., Sullivan, 1980; Steidtmann et al., 1983; Roehler, 1992b; Pietras et al., 2003; Smith et al., 2014b, 2015).

The overall change from balanced-fill to underfilled conditions that occurred at the boundary of the Tipton and Wilkins Peak members is thought to have been driven primary by either changing aridity (Bradley 1929, 1973; Bradley and Eugster, 1969; Eugster and Surdam, 1973; Eugster and Hardie, 1975; Smoot, 1983; Roehler, 1993) or tectonic isolation of the basin (Pietras and Carroll, 2006). A tectonic cause is also implied by its coincidence with structural movements along the Wind River Range to the north and the Uinta Thrust System to the south, as inferred from rapid lithologic and thickness changes along the basin margins (Steidtmann et al., 1983; Roehler, 1993; Bradley, 1995; Pietras et al., 2003).

Sedimentary deformation features are concrete lithological evidence of syndepositional earthquakes emanating from fault systems controlling basin attributes and they support the notion that

the regional shift in the paleohydrological conditions of Lake Gosiute was tectonically induced through the diversion of the regional drainage pattern and the isolation of the basin, rather than due to paleoclimate changes. Furthermore, the timing of these earthquakes, at ~51.5 Ma, coincides with structural movements along the western margin of the basin inferred from the presence of steeply inclined beds, offsets, fault-propagation folds, various mass-transport deposits, and seismites (e.g., Dorr and Gingerich, 1980; Coogan, 1992; Törő and Pratt, 2015a). Together, these events point to an active period in the tectonic history of both the Sevier Fold and Thrust Belt and the Laramide Orogeny that influenced the paleohydrological conditions of the region.

5. Conclusions

Syn depositional tectonic activity during the formation of the Eocene Green River Formation in the Bridger Basin has been inferred previously by the lateral and vertical distribution of lacustrine facies, abrupt changes in stratigraphic thickness, and the shift of lake depocentres. In deposits of paleolake Gosiute it is now also demonstrated by the presence of sedimentary deformation features in outcrops and cores. Deformation caused by chaotic external stresses is manifested by a complex suite of millimetre- to centimetre-scale brittle–ductile and liquefaction features in the form of folds, load structures, pinch-and-swell structures, microfaults, breccias, and sedimentary dikes, the last involving sediment shrinkage and injection. These occur clustered in up to 22 distinct intervals within profundal to sublittoral carbonate-rich lacustrine deposits. The association of brittle and ductile deformation structures in a single interval reflects the heterogeneous rheological state of individual layers at the time of deformation. Superimposition of structures with different deformation style indicates that the interval was affected by more than one event; a change in response from ductile to brittle records increasing sediment stiffness with time. Trona beds responded to the elevated stresses in a brittle–ductile fashion, as shown by trona breccias associated with folding and faulting of the interbedded carbonate mudstone beds. Deformed intervals are under- and overlain by undeformed units of the same

facies, and can be traced laterally for more than 50 km. These features are confidently interpreted as earthquake-induced, and they compare closely with other examples of seismites in low-energy facies.

These seismites clearly have a regional significance, because they coincide with the boundary between two distinct phases of Eocene Lake Gosiute, a balanced-filled stage (Tipton) followed by an underfilled (Wilkins Peak) stage. Thus, they indicate a phase at ~51.5 Ma characterized by high seismic activity that marks rapid, basin-wide changes in lake evolution and, consequently, the stratigraphic architecture of the Greater Green River Basin. The earthquakes were likely related to activity of the basin-bounding Uinta Thrust System, or other faults yet to be delineated within the basin itself. Based on previous studies and empirical relationships between liquefaction and earthquake parameters, these seismites suggest that the earthquakes responsible for their formation had minimum magnitudes between 6.4 and 7.1, and minimum epicentral intensity (MCS) of ~9.4. Seismite-bearing horizons indicate a mean earthquake recurrence period of ~8.1 k.y, but their temporal distribution shows two episodes with enhanced earthquake frequency followed by relative quiescence.

This study demonstrates the utility of stratigraphic frequency and lateral distribution of seismites in rheologically susceptible sediments in low-energy environments with the appropriate preservation potential to help understand the tectonic context and evolution of ancient lake systems.

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Figure 1. Regional geological map of the Greater Green River Basin, southwestern Wyoming, USA (modified after Smith et al., 2008); depocentres after Roehler (1992b); extent of the Rife Bed of the Tipton and lower Wilkins Peak members from Roehler (1993); area of bedded evaporites after Wiig et al. (1995); Eocene faults from Love (1970), Witkind and Grose (1972), Steidtmann et al. (1983), and Bradley (1995). Cross-section X–X' section is shown in Figure 2A. Area marked with red rectangle is shown on Figure 2B.

Figure 2. A: Stratigraphic cross section X–X' of the Green River Formation in the eastern Bridger Basin (modified after Pietras and Carroll, 2006; Smith et al., 2015). A–I – arkosic clastic beds of Culbertson, 1961), Sb – Sheggs bed of Tipton Member, Rb – Rife bed of Tipton Member. Dated tuffs: F – Firehole tuff, 51.41 ± 0.21 Ma; S – Sheggs tuff: 52.22 ± 0.09 Ma (Smith et al., 2008, 2010). B: Map of southeastern Bridger Basin showing locations of studied cores and, outcrops, and major faults. SJC – Slippery Jim Canyon, MFC – Middle Firehole Canyon, FC – Firehole Canyon, SC – Sage Creek, RC – Red Creek, WM1 – U.S. ERDA White Mountain #1 core, BF 1 – U.S. ERDA Blacks Fork #1 core, CCR 1 – U.S. DOE/LETC Currant Creek Ridge #1 core. C: Panoramic view of the Middle Firehole Canyon area, showing member boundaries with blacked dashed lines and the Tipton–Wilkins Peak boundary with red dashed lines. Tw – Wasatch Formation (main body), TwN – Niland Tongue, Tg – Green River Formation, Tglu – Luman Tongue, Tgts – Scheggs Bed of Tipton Member, Tgtr – Rife Bed of Tipton Member, Tgw – Wilkins Peak Member, Tgla – Laney Member. The hill is 210 m above valley floor. Location of A and B is on Figure 1.

Figure 3. Measured stratigraphic outcrop sections of the uppermost Tipton Member and lowermost Wilkins Peak Member in southeastern Bridger Basin, Wyoming, showing the distribution of sedimentary deformation features in distinct horizons (red star). Numbers next to the Red Creek section depict the dominant strike orientation of sedimentary dikes.

Figure 4. Measured stratigraphic sections of cores from the southeastern Bridger Basin, showing the distribution of sedimentary deformation features (red star). Cores were logged at the U.S.G.S. core repository in Denver, Colorado. Core locations: WM1 – section 7, T. 18 N., R. 106 W / $41.550856, -109.418522$; BF1 – section 24, T. 16 N., R. 108 W / $41.356506, -109.524952$; CCR1 – section 14, T. 14 N., R. 107 W. / $41.195948, -109.436232$.

Figure 5. Folding and convolute lamination. A: Folded oil shale with interbedded massive silty mudstone (CCR1). B: Folded oil shale and interbedded massive mudstone (WM1). C: Recumbent folded oil shale (MFC). D: Folding in interlaminated silty mudstone intervals, with microfaults and structureless disrupted mudstone interbeds (MFC). E: Irregularly folded silt layers interbedded with disrupted beds (FC). F: Thin-section photomicrograph showing a convoluted muddy layer overlain by gently folded siltstone laminae (SC).

Figure 6. Load structures. A: Structureless disrupted bed with internal irregular folding loaded into the underlying laminated and microfaulted beds (SJC). B: Disrupted bed with folded and brecciated internal structure and a load structure at its base (SJC). C: Thin-section photomicrograph showing loading of massive muddy interval into underlying laminated and microfaulted laminae (SJC). D: Thin-section photomicrograph showing a load structure filled with disrupted, brecciated material of laminated deposits (SJC). E: Load structure associated with Type 3 sedimentary dike below (black arrow) (SJC). F: Pseudonodules in laminated sandstone (SC).

Figure 7. Hybrid brittle–ductile structures. A: Crumpled sandy siltstone interval with folded, microfaulted laminae (WM1). B: Sandy siltstone interval showing dominantly brittle deformation and isolated Type 4 sedimentary dikes, some of which protrude upward (CCR1). C: Complex deformation feature with large- and small-scale folding and brecciation, sandwiched by less deformed and microfaulted intervals (MFC). D–H: Examples of hybrid structures from SJC. D: Folded and faulted intercalated calcareous siltstone-mudstone with disrupted and brecciated muddy interbeds. E: Area outlined in B, with microfaults affecting both the laminites and the structureless muddy interbed. F: Thin-section photomicrograph showing microfaults associated with highly convoluted laminites. G: Slab showing a basal convoluted interval overlain by folded and faulted laminites. H: Load structure composed of structureless muddy layer associated with microfaults and millimetre-scale sedimentary dikes (Type 4).

Figure 8. Pinch-and-swell structures. A: Pinch-and-swell structures with microfaults overlain by laminated oil shale (BF1). B: Pinch-and-swell structures with microfaults overlain by microfaulted and folded oil shale (BF1).

Figure 9. Microfaults. A: Variably tilted normal and reverse faults (white arrows) in laminated oil shale (CCR1). B: Subparallel set of normal microfaults, over- and underlain by folded laminae with load structures (SJC). C: Thin-section photomicrograph showing a normal microfault in a silt- and sand-rich interval, associated with folded laminae and isolated and upward-protruding Type 4 sedimentary dikes (black arrows) (MFC). D: Interbedded siltstone-and-mudstone interval showing an array of reverse microfaults, associated with a mud-filled sedimentary dike (Type 2?, white arrow) and brecciation (to the left) (FC). E: Two faulted zones (white arrow) showing reverse displacement, associated with the fracturing of sand-rich intervals and the formation of mud-filled fractures. Faulted layers change gradually into a disrupted and brecciated interval to the right (FC).

Figure 10. Breccias. A: Brecciated interval of silty mudstone (SJC). B: Fragmented siltstone bed and siltstone clasts within structureless sandstone beds, associated with sand-filled sedimentary dikes and a flame structure (middle) (SC). C: Subhorizontal breccia bodies within disrupted planar laminated/wave-rippled sandstone bed (SC). D: Tracing of C, showing brecciated intervals (grey) and clasts (dark grey). E: Oblique brecciated zone with disrupted sandy siltstone host interval (SC). F: Tracing of E, showing the brecciated zone (grey), the adjacent disrupted zone in the host sediment (outlined with dotted line), and dislocated clasts (dark grey).

Figure 11. Oil shale breccias. A: Brecciated oil shale in homogeneous, organic-rich mudstone (BF1). B: Two brecciated oil shale intervals (white arrows) divided by laminated host sediment (BF1). C: Brecciated oil shale with a laminated oil shale clast (white arrow) from the underlying laminated host strata (BF1). D: Breccia in relatively organic-poor oil shale, with organic-rich clasts (BF1). E: Oil shale breccia with a fragmented mudstone interbed in the middle (white arrow) (WM1). F: Oil shale breccia (base) overlain by massive and laminated oil shale (CCR1).

Figure 12. Type 1 sedimentary dikes. A: Isolated, bulbous, massive mud-filled dike in laminated oil shale (BF1). B: Upward-protruding sedimentary dike (white arrow) in laminated oil shale and filled with massive mudstone (CCR1). C: Two sheet-like dikes (white and black arrows) filled with silty mudstone, hosted by microfaulted laminated oil shale (BF1).

Figure 13. Type 2 sedimentary dikes. A: Metre-long sedimentary dike following a jagged path penetrating through several siltstone beds (MFC). B: Two cross-cutting sets of metre-scale dikes in laminated carbonate mudstone (RC). C: Downward-tapering dike showing a step-like lateral shift (RC). D: Close-up of area outlined in A, showing normal displacement and folding of a siltstone bed penetrated by a dike. E: Downward-tapering dike crosscutting irregular and deformed silty sand intervals with type 4 dikes, and overlain by a wave-rippled sandstone bed (RC). F: Thin-section photomicrograph of a dike filled with plastically deformed mixture of mud, silt and fragmented material (RC). G: Thin-section photomicrograph of a dike showing plastic deformation of multiple generations of silt- and sand-rich infills organized into sedimentary dikes (RC). Small-scale features, similar in morphology to type 4 dikes, penetrate the host laminated mudstone.

Figure 14. Type 3 sedimentary dikes at RC. A: Two dikes with a mudstone 'plug' (white arrows) and a brecciated source bed with irregular thickness (black arrows). B: Area outlined in A, showing a folded dike with brecciated lower part and massive mudstone above, and second generation infill (white arrow). C: Brecciated infill forming a sill from the vertical dike (boundary marked by dotted line). D: Breccia-filled dike (boundary marked by dotted line) overlain by a brecciated source interval with irregular thickness (black arrow).

Figure 15. Type 3 sedimentary dikes at RC. A: Isolated and tilted dike filled with brecciated mudstone. B: Slabbed surface showing brecciated and in-situ microbial structures. C: Thin-section photomicrograph showing two generations of infill: a brecciated silty infill on the sides, a second infill with fragmented microbial structures floating in a muddy matrix.

Figure 16. Type 4 sedimentary dikes. A: Variably tilted, downward-tapering dikes at the base of a massive silty mudstone in laminated oil shale, with microfaults below (white arrow) (BF1). B: Upward- and downward-tapering silt-filled dikes in laminated silty mudstone (WM1). C: Densely spaced network of silt-filled dikes in interbedded parallel-laminated siltstone and mudstone interval (RC). D: Upward- and downward-tapering dikes in wavy-laminated sandstone (MFC). E: Thin-section photomicrograph showing isolated and upward- and downward-tapering dikes penetrating silty and muddy interval (MFC). F: Tilted and folded carbonate mud-filled dikes in wavy-

laminated sandstone, overlain by a disrupted and folded interval (FC). G: Thin-section photomicrograph showing a disrupted mud-filled dike with fragmented infill in laminated sandstone (FC).

Figure 17. Type 5 sedimentary dikes. A, B: Dense network of spindle-shaped and triradiate dikes at the base of a siltstone bed (RC). C–I: Shapes of dikes on bedding planes, traced from photographs at MFC (C) and RC (D–F). C: Irregularly lenticular and triradiate dikes. D: Sub-parallel linear and lenticular dikes. E, F: Randomly oriented irregularly spindle-shaped dikes with complex, composite forms. G: Lenticular and triradiate dikes with preferred orientation. Orientation of the dikes (white dashed lines) is slightly different in successive layers shown by differing shades of grey. H: Linear and parallel lenticular dikes. I: Randomly oriented linear and lenticular dikes.

Figure 18. Deformed tuffs. A: Tuff with disturbed bedding, irregular base and folded top, and fluid-escape cusp (white arrow) (BF1). B: Disturbed and massive tuff with irregular tuff lenses above (WM1). C: Two massive tuff beds with irregular top and base, and the upper one shows tilted sedimentary dikes penetrating downwards from its base (WM1). D: Isolated tilted and sub-horizontal tuff-filled sedimentary dikes at the base of a tuff layer (BF1). E: Isolated and tilted tuff-filled sedimentary dikes at the base of a tuff layer (BF1).

Figure 19. Deformed trona in BF1 core. A: Fracture filled with white fibrous trona overlain by brecciated coarse crystalline trona with plastically deformed mudstone fragments. The deformed interval is underlain by oil shale breccia and overlain by microfaulted, laminated carbonate mudstone. B: Folded laminated carbonate mudstone within a brecciated trona interval showing angular trona clast and a microfault in the lower part (white arrow).

Figure 20. Simplified sketch, description, and deformation mechanism of earthquake-induced sedimentary deformation features (seismites) observed in profundal to sublittoral deposits of the Tipton–Wilkins Peak transition in the southeastern Bridger Basin, Wyoming.

Figure 21. Map showing the number of seismites in the Tipton Member (blue) and Wilkins Peak Member (pink) for each section and core location.

Figure 22. Stratigraphic distribution of seismites in cores indicating clustering of seismic events, with two active periods followed by relatively quiescent intervals. Correlation of distinct deformed intervals is marked with black dotted lines. The Tipton–Wilkins Peak boundary is marked with red dashed line. Light-grey shading indicates proposed correlation of the clustered intervals. Based on lithological correlation (Walker, 2008), the lowermost seismites in WM1 (dark-grey shading) may be stratigraphically older than those in the other cores.

Figure 23. A: Distribution of temporal intervals for each studied core and clustering (shaded areas) of seismic events, calculated from the distance of seismites and average depositional rate of 110 mm/k.y. B: Three-dimensional graph showing the distribution of temporal intervals for all three cores and the noticeable clustering.

Sedimentary Deformation Structure	Description	Deformation Mechanism
<p>Folds/convolute lamination</p>	irregular, mm- to cm-scale folding with open to tight geometries in oil shale and laminated silty mudstone	chaotic compressive external stresses, and squeezing acting on semi-consolidated or liquified sediment
<p>Load structures</p>	irregular, mm- to cm-scale load structures at the base of massive or disrupted mudstone interbeds in folded laminated silty mudstone	compressive stress; dominantly squeezing of disrupted mudstone into underlying semi-consolidated or liquified sediment
<p>Hybrid (brittle-ductile) features</p>	complex association of folds, load structures, microfaults, and Types 2, 4, and 5 sedimentary dikes in laminated silty mudstone	chaotic tensile and compressive external stresses, coupled with squeezing, dewatering, shrinkage, and injection, in semi-consolidated or liquified sediment
<p>Pinch-and-swell structures</p>	mm- to cm-scale thickening and thinning bundles of laminae in oil shale, associated with normal microfaults	tensile external stress causing stretching and, locally, brittle failure in semi-consolidated or liquified sediment
<p>Microfaults</p>	mm- to cm-scale faults, with varying style, offset, and dip angle in laminated oil shale or silty mudstone, locally associated with Types 2, 4, and 5 sedimentary dikes	chaotic compressive and tensile external stresses causing brittle failure of semi-consolidated to consolidated sediment
<p>Breccias</p>	dm-scale disrupted beds, breccia layers, and irregular breccia bodies with a massive muddy or sandy matrix in folded laminated silty and sandy mudstone	chaotic tensile, compressive and shear stresses causing brittle failure and disruption of semi-consolidated beds, and liquidization of less consolidated beds, superimposed by remobilization and injection
<p>Oil shale breccias</p>	dm-scale stratiform brecciated oil shale with massive organic-rich matrix, variably organic-rich intraclasts, and sharp upper and lower boundaries	external shear stresses causing elevated pore pressure, liquefaction, in-situ brittle failure and disruption of semi-consolidated oil shale, or mass-transport processes

Sedimentary Deformation Structure	Description	Deformation Mechanism
<p>Type 1 sedimentary dikes</p>	cm- to dm-scale isolated, bulbous or spindle- shaped dikes filled with massive mudstone in laminated oil shale	elevated pore pressure causing dewatering and shrinkage of the semi-consolidated oil shale, associated with the segregation, and injection of carbonate mud
<p>Type 2 sedimentary dikes</p>	dm- to m-scale, tabular or downward-tapering dikes filled with a mixture of carbonate and silt in silty mudstone, associated with folds, microfaults, and Types 4 and 5 sedimentary dikes	elevated pore pressure causing fluidization of mud and silt, and injection into fissures formed by brittle failure and hydrofracturing of the semi-lithified host strata
<p>Type 3 sedimentary dikes</p>	dm-scale, folded, downward tapering dikes filled with brecciated material with carbonate mud matrix, associated with Types 2 and 4 sedimentary dikes	external shear stresses causing disruption of semi-consolidated beds, fluidization, and subsequent infill of subaqueous fissures, followed by remobilization, injection, and brecciation
<p>Type 4 sedimentary dikes</p>	downward- and upward-tapering, and isolated mm- to cm-scale dikes in oil shale and intercalated carbonate mudstone, calcareous siltstone, and sandstone, filled with a mixture of mud, silt and sand	elevated pore pressure causing fluidization of mud and silt, and injection into fissures formed by brittle failure and hydrofracturing or dewatering and shrinkage of the muddy semi-lithified host
<p>Type 5 sedimentary dikes</p>	cm-scale dikes that on bedding surfaces are doubly tapering, straight to curvilinear, or branching, and filled with a mixture of mud and silt
<p>Deformed trona</p>	brecciated beds of angular trona and deformed mudstone clasts, hosted by laminated carbonate mudstone showing hybrid brittle-ductile deformation	variably directed compressive, tensile, and shear stresses causing in-situ disruption and short distance translation of trona beds
<p>Deformed tuff</p>	plastically deformed tuff beds, with irregular thickness, fluid-escape cusps, oil shale clasts, and tuff-filled, downward- tapering or isolated Type 4-style dikes at the base	elevated pore pressure causing liquefaction, with fluid escape and injection into fissures formed by dewatering and shrinkage of the host oil shale

SJC Slippery Jim Canyon
 MFC Middle Firehole Canyon
 FC Firehole Canyon
 SC Sage Creek
 RC Red Creek
 WM 1 White Mountain #1
 BF 1 Blacks Fork #1
 CCR1 Currant Creek Ridge #1

Table 1. Calculated sedimentation rates in mm/k.y. for WM1 and CCR1 cores.

Core	Thickness (m)	Time (k.y.) ^a	mm/k.y.
WM1	74.6	810	92
CCR1	99.3	810	122

Calculations are based on dated tuffs of Smith et al. (2008, 2010), and the stratigraphic thickness between the dated tuffs using published data of Pietras (2003) and Walker (2008).

^aTime span of dated tuffs without analytical errors.

Table 2. Calculated recurrence periods using the total number of seismites.

Location	Studied thickness (m)	Time span (k.y.) ^a	Number of seismites	Recurrence period1 (k.y.) ^a	Recurrence period2 (k.y.) ^b
SJC	5	45.5	5	9.1	
MFC	5	45.5	7	6.5	
FC	5	45.5	5	9.1	
SC	7	63.6	6	7.5	
RC	26	236.4	15	15.8	
WM1	15 m	136.4	18	8.1 k.y.	9.7 k.y.
BF1	15 m	136.4	22	6.6 k.y.	
CCR1	15 m	136.4	19	7.7 k.y.	6.9 k.y.

^aBased on average sedimentation rate of 110 mm/k.y. (Smith et al., 2008).

^bBased on calculated sedimentation rates (Table 1).

Table 3. Calculated recurrence periods in the three studied cores using temporal intervals.

WM1 core				
Seismites	Seismites thickness (cm)	Interval thickness (cm) ^a	Interval time1 (k.y.) ^b	Interval time2 (k.y.) ^c
1	6			
2	4	15.9	1.4	1.7
3	8	18.3	1.7	2.0
4	2	12.2	1.1	1.3
5	6	36.6	3.3	4.0
6	15	189.0	17.2	20.5
7	8	39.6	3.6	4.3
8	7	42.7	3.9	4.6
9	2	48.8	4.4	5.3
10	3	103.6	9.4	11.3
11	15	213.4	19.4	23.2
12	17	198.1	18.0	21.5
13	18	24.4	2.2	2.7
14	36	42.7	3.9	4.6
15	4	70.1	6.4	7.6
16	2	121.9	11.1	13.3
17	15	76.2	6.9	8.3
18	4	143.3	13.0	15.6
		AVERAGE	7.5	8.9
		S.D.	6.1	7.3
BF1 core				
Seismites	Seismites thickness (cm)	Interval thickness (cm) ^a	Interval time1 (k.y.) ^b	
1	14			
2	7	45.7	4.2	
3	2	6.0	0.5	
4	18	48.8	4.4	
5	5	6.0	0.5	
6	2	57.9	5.3	
7	3	48.8	4.4	
8	2	85.3	7.8	
9	24	125.0	11.4	
10	45	155.4	14.1	
11	20	143.3	13.0	
12	75	70.1	6.4	
13	15	15.2	1.4	
14	70	76.2	6.9	
15	54	57.9	5.3	
16	6	9.1	0.8	
17	10	48.8	4.4	
18	1	42.7	3.9	
19	2	27.4	2.5	
20	6	9.1	0.8	
21	19	140.2	12.7	
22	4	118.9	10.8	
		AVERAGE	5.8	
		S.D.	4.4	
CC1 core				
Seismites	Seismites thickness	Interval	Interval time1 (k.y.) ^b	Interval time2 (k.y.) ^c

	(cm)	thickness (cm) ^a		
1	4			
2	2	27.4	2.5	2.2
3	2	6.1	0.6	0.5
4	5	106.7	9.7	8.7
5	2	21.3	1.9	1.7
6	3	9.1	0.8	0.7
7	10	97.5	8.9	8.0
8	12	149.4	13.6	12.2
9	6	9.1	0.8	0.7
10	7	213.4	19.4	17.5
11	4	18.3	1.7	1.5
12	5	30.5	2.8	2.5
13	15	21.3	1.9	1.7
14	18	24.4	2.2	2.0
15	2	27.4	2.5	2.2
16	4	36.6	3.3	3.0
17	17	307.8	28.0	25.2
18	6	45.72	4.2	3.7
19	4	61	5.5	5.0
		AVERAGE	6.1	5.5
		S.D.	7.4	6.7

^aInterval thickness measured from the top of the deformed layer to the top of the next overlying one.

^bInterval time calculated from the interval thickness and average sedimentation rate of 110 mm/k.y.

^cInterval time based on calculated sedimentation rates in the case of WM1 and CCR1 core (Table1).

S.D. – standard deviation