A dinoflagellate cyst zonation of the Cenomanian and Turonian (Upper Cretaceous) in the Western Interior, United States

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ABSTRACT
This study documents the detailed palynology of the Bridge Creek Member, Greenhorn Limestone Formation, at the Rock Canyon anticline outcrop, west of Pueblo, Colorado, USA. The section is the formally ratified Global Boundary Stratotype Section and Point (GSSP) for the base of Turonian Stage and corresponding Lower Turonian Substage, and the only proposed GSSP for the base of the Middle Turonian Substage. The authors previously published palynological data from the underlying Cenomanian strata in the nearby US Geological Survey (USGS) Portland-1 core, and from cores and an outcrop of Cenomanian, Turonian and earliest Coniacian age in SW Texas, are reviewed for biostratigraphy. Dinoflagellate cyst ranges and events in both areas are calibrated against other published bio- and chemostratigraphical (carbon isotope) data and an astronomical age model. A regional dinoflagellate cyst zonation is proposed, adapted from a Central and Northern European scheme, comprising three zones and four subzones. Comparison with vintage published palynological data from other states, including Arizona, Kansas, Utah, Wyoming and Montana, indicates that the zonation is probably applicable to marine sections throughout the Cretaceous Western Interior Seaway of the USA. The new palynological data from around the Lower–Middle Turonian boundary at Pueblo support its proposal as a GSSP.

1. Introduction
1.1. Background and objectives
Dinoflagellate cysts are widely used in biostratigraphical studies of the Upper Cretaceous Series. They have been used to guide conventional hydrocarbon exploration and development drilling in many regions, including Wyoming, USA (the Frontier Formation and Greenhorn Formation; Burgess 1971) and Northern Europe, in both the northerly argillaceous Shetland Group facies (Costa & Davey 1992) and southerly calcareous Chalk Group facies (Schsieler & Wilson 1993). More recently, they have been proposed for use in guiding unconventional shale oil/shale gas reservoir drilling of the Eagle Ford Group in Texas, USA (Eldrett et al. 2015a; Dodsworth 2016). It has been known for some time that several dinoflagellate cysts have widespread range bases and tops in the Cenomanian and Turonian stages and that they are therefore of utility in identifying their substages (Clarke & Verdier 1967; Burgess 1971; Foucher 1979, 1980, 1981, 1983; Williams & Bujak 1985; Williams et al. 2004).

The overall objectives of the Subcommission on Cretaceous Stratigraphy (SCS) which fit within the International Union of Geological Sciences’ policy are: to establish a standard global stratigraphical subdivision and nomenclature for the Cretaceous; to produce a stratigraphical table displaying agreed subdivisions to stage and substage level, marking boundaries that are defined by a Global Boundary Stratotype Section and Point (GSSP; Remane et al. 1996; http://cretaceous.stratigraphy.org). Following the formal ratification of GSSPs for some of the Upper Cretaceous stage boundaries, i.e. the bases of the Cenomanian Stage (Kennedy et al. 2004), Turonian Stage (Kennedy et al. 2005), Santonian Stage (Lamolda et al. 2014) and Maastrichtian Stage (Odin & Lamaurelle 2001), it is appropriate to undertake detailed palynological calibrations of these global reference sections. Additionally, palynological studies of proposed GSSPs, such as a report on dinoflagellate cysts across a candidate base of the Coniacian Stage GSSP in Poland (Olde et al. 2016), strengthen the case for their selection by the SCS.

Upper Cretaceous stages are divided into lower, middle and upper substages. The base of a lower substage is defined by the same GSSP as the corresponding base of stage boundary. However, middle and upper substage boundaries have, to date, received less attention and none are yet formally ratified. One of the best documented is the proposed GSSP for the base of the Middle Turonian Substage (Bengtson et al. 1996; Kennedy et al. 2000), at the same locality as the formally ratified GSSP for the base of the Turonian Stage/Lower Turonian Substage (Kennedy et al. 2005), Rock Canyon anticline outcrop, west of Pueblo, Colorado, USA (abbreviated to ‘Pueblo’ in the text; Figures 1–3; Plates 1 & 2) and the nearby USGS core...
The Upper Cenomanian, Lower Turonian and Middle Turonian substages are represented by the Lincoln Shale Member, Hartland Shale Member and Bridge Creek Member of the Greenhorn Limestone Formation in central Colorado (Figures 4 and 5). Previous palynological investigations of the Upper Cenomanian to Lower Turonian interval are from Pueblo: (1) Courtinat (1993), 21 spot samples from the lower Bridge Creek Member (S. gracile to lower W. coloradoense Ammonite Zones), focusing on possible relationships between palynofacies and lithology; (2) Li and Habib (1996), 14 spot samples from the same interval, focusing on palynofacies and the ratio of chorate to proximochorate-proximate dinoflagellate cysts; (3) Dodsworth (2000), 53 samples from the lower to middle Bridge Creek Member (S. gracile to lower M. nodosoides Ammonite Zones), focusing on biostratigraphy and palaeoenvironments, published in support of the then proposed (and subsequently ratified) base of the Turonian GSSP; (4) Harris and Tocher (2003), 42 samples from a comparable interval (S. gracile to basal M. nodosoides Ammonite Zones), focusing on regional correlation with coeval intervals from three other central KWS sections: Bunker Hill in Kansas, ca. 600 km to the east; Wahweap Wash in Utah, ca. 600 km to the west; and Blue Point in Arizona, ca. 730 km to the southwest. None of these previous palynological studies covers the higher part of the outcrop at Pueblo, including the Lower–Middle Turonian Substage boundary (upper M. nodosoides to C. woollgari Ammonite Zones). There are also no published palynological studies of marine sections form the Upper Turonian Substage and Coniacian Stage in Colorado. Previous palynological investigations of the Upper Cenomanian, Turonian and lowermost Coniacian interval of the upper Eagle Ford Group and overlying Austin Chalk Formation in SW Texas are by Eldrett et al. (2014, 2015a, 2015b, 2017) and Dodsworth (2016).

This paper documents 26 samples from the higher part of the Pueblo outcrop, Beds 110 to 125, which for the first time provide palynological data from strata that span the proposed base of the Middle Turonian Substage GSSP (Bed 120). It also provides a re-issuance of data from the underlying 53 samples previously reported by Dodsworth (2000), Bed 63 (base of Bridge Creek Member) to Bed 109, with updated dinoflagellate cyst nomenclature, following the recommendations of Williams et al. (2017), including taxa published since the year 2000, e.g. Senonisphaera turonica (formerly Canningia sp. A of Marshall & Batten 1988 and Senonisphaera rotundata alveolata) of Pearce et al. 2003, 2009 and Prince et al. 2008). Due to space constraints, some taxa were grouped at generic level on the Pueblo range chart in Dodsworth (2000), e.g. Heterosphaeridium, Isabelidinium/ Eurydinium and Subtilisphaera, but were discussed in the corresponding text at species level. Here, they are plotted at species level in a larger supplemental range chart. More detail is also provided on the supplemental range chart for terrigenous pollen and spores and for reworked taxa.
The authors’ palynological data from the Portland-1 core, which were previously published and discussed with emphasis on palaeoenvironments (Eldrett et al. 2017), are reviewed here with respect to Cenomanian biostratigraphy. Our high-resolution palynological studies of the Cenomanian, Turonian and lowermost Coniacian deposits in SW Texas are also reviewed and correlation with the Pueblo and Portland-1 bioevents is discussed. The data for Portland-1 core and SW Texas cores Iona-1, Innes-1 and ‘Well X’ (Figure 2) are archived at the data repository PANGEA. The Digital Object Identifier DOI of the dataset is https://doi.org/10.1594/PANGAEA.877411. Eldrett et al. (2015a) used uranium–lead (U–Pb) dates from individual zircons from 10 bentonite layers in Iona-1 to provide geochronological constraints for astronomical (obliquity, eccentricity and average spectral misfit) age models for individual events, including 12 dinoflagellate cyst bioevents. Here, ages for a further 23 organic-walled phytoplankton bioevents in Iona-1 have been derived using the obliquity model. The method has also been applied to Portland-1 (Eldrett et al. 2017). A formal dinoflagellate cyst zonation of the Cenomanian, Turonian and lowermost Coniacian in the Western Interior of the US, calibrated to these ages, is proposed here for the first time and applied to the sections documented.

Figure 2. Sketch maps of locations. (a) The area around the Pueblo Reservoir (based on Scott 1964), including the Global Stratotype Standard Section on the north side of the reservoir, and the points for the base of the Turonian Stage, Bed 86, and the proposed Middle Turonian Substage, Bed 120 (Kennedy et al. 2000, 2005). Grey shading shows outcrop of Greenhorn Limestone Formation. (b) SW Texas map showing outcrop and core locations (black stars): Lozier Canyon outcrop (29°53.37′N, 101°48.22′W); Innes-1 core (29°49.21′N, 101°37.34′W); Iona-1 core (29°13.51′N, 100°44.49′W); Well ‘X’ core. Grey shading shows outcrop of Eagle Ford Group (from Eldrett et al. 2017).
1.2. Geological setting

The Late Cretaceous Epoch was characterized by a sustained warm climate resulting in high eustatic sea levels and numerous epicontinental seaways, including the KWIS. Several major global perturbations in the carbon cycle occurred, termed ‘oceanic anoxic events’ (OAE), the most prominent spanning at the Cenomanian–Turonian transition and named OAE-2 (Schlanger & Jenkins 1976; Schlanger et al. 1987). This interval in particular was marked by a globally recognized positive carbon isotope ($\delta^{13}C$) excursion, reflecting the widespread removal of $\delta^{12}C$-enriched organic matter in marine sediments under global anoxic conditions (Jenkyns 2010 and references therein). However, the deposition of organic-rich, fine-grained sediments varied both temporally and spatially, being modulated and ultimately dependent on local and regional processes (basin restriction, water stratification, bottom currents, sediment input) in addition to global phenomena (sea level change, orbital forcing; Eldrett et al. 2014, 2015b, 2017; Minisini et al. 2017). Less prominent maxima in the Cenomanian and Turonian $\delta^{13}C$ curve in the KWIS, e.g. a Middle Cenomanian Event (MCE), can also be correlated inter-regionally (Figure 6), particularly with England and Japan (Jarvis et al. 2006; Uramoto et al. 2013; Joo & Sageman 2014; Eldrett et al. 2015a).

The KWIS submerged North America along a north–south axis (Figure 1), at times mixing Boreal Realm waters from the north (present-day Canada) with Tethyan Realm waters from the south (Gulf of Mexico and proto-Atlantic Ocean) (Kauffman 1977, 1984a, 1985; Eicher & Diner 1985; Hay et al. 1993; Kauffman & Caldwell 1993; Slingerland et al. 1996; Elderbak & Leckie 2016; Eldrett et al. 2017; Lowery et al. 2018). In this paper, we refer to northern KWIS, including marine deposits in Canada, central KWIS including the USA to the north of Texas, and southern KWIS (Texas).

The Dakota Sandstone Formation (Albian–earliest Cenomanian; Section 4.1) is one of the oldest Cretaceous sedimentary units in central Colorado (Gustason & Kauffman 1985), recording both nearshore marine and non-marine deposition during onset ‘Greenhorn’ marine transgression (Kauffman 1977). In Portland-1, it contains long-ranging terrigenous spores and pollen in the lower part, 211.61 to 211.53 m, and dinoflagellate cysts associated with freshwater algae from 209.59 to 205.11 m (Eldrett et al. 2017). At 204.39 m, it is overlain unconformably by the marine Graneros Shale Formation of Early to Middle Cenomanian age, including the Thatcher Limestone Member. Dinoflagellate cysts remain relatively abundant and diverse throughout the Graneros Shale (Figure 4).

In Texas, the Buda Limestone Formation was initially deposited during the Early Cenomanian but is overlain unconformably by the organic-rich (ca. 1 to >6% total organic carbon) lower Eagle Ford Group. The Buda Limestone...
comprises highly diverse and low-abundance dinoflagellate cyst assemblages, dominated by gonyaulacoid taxa, ‘G-cysts’ (Cornell 1997; Eldrett et al. 2017), probably reflecting a low-nutrient, oligotrophic setting (cf. Harland 1988; Sluijs et al. 2005). By contrast, the lower Eagle Ford is generally characterized by a decline in dinoflagellate cyst species diversity, the palynological assemblages comprising high absolute abundance of green algal prasinophyte phycomata (Tyson 1995), peridinioid taxa, ‘P-cysts’, being the major component of the dinoflagellate cyst community (Eldrett et al. 2015b, 2017; Dodsworth 2016), indicative of eutrophic and stratified water column conditions (cf. Prauss 2007; Sluijs et al. 2005).

During Middle Cenomanian to Middle Turonian times, the marine Greenhorn Limestone Formation was deposited in central Colorado, comprising the interbedded restricted to open marine Lincoln Shale Member (Sageman & Johnson 1985), the mainly restricted marine Hartland Shale Member (Sageman 1985) and the more open-marine Bridge Creek Member. A δ¹⁸O carbonate curve from Pueblo published by Pratt (1985, Figure 11), shows a broad up-section positive shift in values between the base of the Lincoln Shale and base of the Bridge Creek, interpreted as reflecting an increase in salinity to normal marine levels in the central KWIS during the peak Greenhorn transgression. Dinoflagellate cysts comparable to those informally described from Tarfaya, Morocco (Figure 1) by Prauss (2012a, 2012b, 2012c, 2015), Bosedinia cf. sp. 1 and 3 of Prauss (2012b), which had a mainly proto-Atlantic Tethyan Realm abundant distribution during the Late Cretaceous (Eldrett et al. 2017), became established in abundance during deposition of the lower Eagle Ford in SW Texas, below and above the regionally correlatable ‘X-bentonite’ in Iona-1 and Innes-1, and over a shorter stratigraphical range farther north in Portland-1, above the X-bentonite (ca. 173.5 m), in the uppermost Lincoln Shale and Hartland Shale (Figure 4). In both areas their abundance is associated with low-diversity agglutinated benthic foraminiferal assemblages, redox-sensitive trace metal enrichments and organic biomarkers, consistent with deposition under oxygen-deficient conditions (Sun et al. 2016; Eldrett et al. 2017; Minisini et al. 2017).

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the Bridge Creek limestone and bentonite beds at Pueblo can be correlated across wide areas of the central KWIS (Hattin 1971, 1975, 1985; Sageman et al. 2006). Despite a globally contemporaneous carbon isotope (δ13C) excursion, sediments of the Bridge Creek and upper Eagle Ford were mostly oxygenated during OAE-2 in the KWIS, as indicated by limited or no enrichment in redox-sensitive elements (Meyers 2007; Denne et al. 2014; Eldrett et al. 2014, 2017; Figure 5).

The Cenomanian–Turonian transition was an interval of extreme warmth (Huber et al. 2002; Forster et al. 2007). However, the early stages of OAE-2 were characterized by a short-lived interval of cooler Boreal waters spreading southwards across Europe, termed the ‘Plenus Cold Event’ (PCE) by Gale and Christensen (1996). In mid-latitude sites including Central Europe and central–southern KWIS, Boreal Realm dinoflagellate cysts, such as Cyclonephelium compactum–Cauveridinium membraniphorum (Figure 5) and Isabelidinium magnum, became consistent in occurrence from Late Cenomanian times onwards which has been interpreted as evidence for the PCE in the KWIS as well as in Europe (Eldrett et al. 2014, 2017; Van Helmond et al. 2016). In the central and southern KWIS, this interval is marked by the persistent abundance of benthic fauna (Elderbak & Leckie 2016 and references therein) and termed the ‘benthonic zone’ (Eicher & Worstell 1970) or the ‘benthic oxic zone’ (Keller & Pardo 2004), occurring at Pueblo in beds 68–78 of the Bridge Creek. This interval of improved water column and seafloor oxygenation has also been linked to inflow of a northerly Boreal water-mass that ventilated the seaway during a period of cooler climate conditions (Eldrett et al. 2014, 2017; Van Helmond et al. 2016). However, some workers still consider the PCE to be mainly restricted to Europe (Elderbak & Leckie 2016).

A negative δ18O shift within the lower Bridge Creek, Beds 78–85 at Pueblo, correlates with relatively organic-rich deposition during OAE-2 there (Figure 5) and was interpreted by Pratt (1985) to reflect the spread of subsaline surface waters into the KWIS during their deposition. This interpretation was supported by Leckie (1985), who reported anomalous large proportions of biserial planktonic foraminifera (Heterohelix spp.) in the beds, and Dodsworth (2000), who reported significantly larger numbers of terrigenous palynomorphs, mainly gymnosperm bisaccate pollen, than in adjacent strata (Figure 5). The rapid rise in heterohelicids (surface-dweller opportunist foraminifera) has since been recorded at a correLate stratigraphical level across the KWIS, including Texas (Donovan & Staerker 2010; Donovan et al. 2012; Lowery et al. 2014; Eldrett et al. 2015a), and in Mexico (Núñez-Useche et al. 2016). It is attributed to the better adaptation of these organisms to sudden environmental change (temperature, salinity, nutrient level) and/or a decline in water column oxygenation (Keller & Pardo 2004; Caron et al. 2006; Reolid et al. 2015) and/or is associated with a brief incursion of a dysoxic proto-Atlantic water-mass into the KWIS (Eldrett et al. 2017). In the central KWIS, a paucity of ammonites and bivalve molluscs from this interval has been correlated with a global faunal mass extinction (Kauffman 1984b; Elder 1987, 1989; Harries & Kauffman 1990; Harries 1993).

In terms of ammonite provincialism, Cobban (1984) reported Cenomanian–Turonian Boreal genera such as Dunveganoceras and Scaphites as having a typically northern

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Figure 5. Pueblo outcrop. Legend same as Figures 3 and 4, except for (k), Ccm is Cyclonephelium compactum–Cauveridinium membraniphorum c.p.g.
KWIS distribution while genera such as Vascoceras, Nigericeras, Romaniceras and Thomasites, which are of value for correlation with Tethyan sequences, are mostly confined to the southernmost part of the KWIS. However, Tethyan ammonite faunas were able to migrate northwards into the central KWIS during intervals of improved water column oxygenation (Batt 1993) and are recorded from above the Hartland Shale – Bridge Creek boundary at Pueblo (Kauffman 1984a). North to south migration also occurred; Kennedy et al. (1987) noted the presence of Boreal ammonite faunas as far south as SW Texas in intra-Lower Turonian strata (Pseudaspidoceras flexuosum Ammonite Zone; cf. Figure 3).

1.3. The proposed base of the Middle Turonian GSSP

The Turonian Working Group of the SCS held meetings during the Second International Symposium on Cretaceous Stage Boundaries in Brussels, 8–16 September 1995. A report of the conclusions of the Working Group was provided by Bengtson et al. (1996). A GSSP at the base of Bed 86 at Pueblo (Figure 2) was proposed and has subsequently been formally ratified by the SCS as the base of Turonian Stage/Lower Turonian Substage (Kennedy et al. 2005). With respect to the base of Middle Turonian GSSP, Bengtson et al. (1996) proposed, ‘The first occurrence of the ammonite Collignoniceras woollgari (Mantell 1822) in the section at Rock Canyon anticline west of Pueblo, Colorado, exposing the Bridge Creek Member of the Greenhorn Limestone. According to current state of knowledge this level lies within Bed 120 and coincides with the first occurrence of the inoceramid bivalves Mytiloides hercynicus (Petrascheck 1903) and Inoceramus cuvierii J. Sowerby, 1814, in the section.’ Following the Brussels meeting, a ballot of the 39 members of the Working Group was 33 in favour of the above proposal, four against and two abstentions. The negative voters argued that the boundary stratotype should be selected in Europe, where the historical stage stratotypes are, and that the Rock Canyon anticline section is still too poorly known, in particular with respect to microfossil stratigraphy. Kennedy et al. (2000) provided further inoceramid data for the section in support of the proposed substage boundary level. The present paper provides the first palynological data, including dinoflagellate cyst stratigraphy, also in support of the proposal, addressing the criticism of insufficient microfossil data. Detailed information on precise location and
access at the proposed base of the Middle Turonian GSSP can be found in Kennedy et al. (2000).

2. Material and methods

Seventy-nine channeled samples (Bridge Creek, BC series) from approximately 12 m of the lower and middle Bridge Creek Member at Pueblo were prepared and analyzed for palynology. The higher 26 samples (BC54 to BC80, Beds 110 to 125) were prepared in the laboratory at the same time as the lower 53 samples (BC01 to BC53, Beds 63 to 109) but had not been analyzed for palynology at the time of publication of Dodsworth (2000). Five grams, or multiples thereof, of crushed material from each rock sample was dissolved in hydrochloric acid (HCl 35%) and hydrofluoric acid (HF 40%) in order to remove carbonate and silicate minerals, respectively. Preparations were sieved with a 10 μm mesh. Palynomorphs and brown and black wood fragments (vitrinite and inertinite) dominate the >10 μm kerogen in the lower part of the succession at Pueblo (S. gracile, lower N. juddii and lower W. coloradoense Ammonite Zones). Some of these preparations contained transparent, ‘cloudy’ amorphous organic matter (AOM) which was removed by a ‘niric wash’, i.e. two minutes of oxidation with nitric acid (70% HNO₃). In the upper N. juddii Ammonite Zone and upper W. coloradoense–M. nodosoides Ammonite Zones, the >10 μm kerogen fraction from the mudstone samples is dominated by dark-colour, clamped AOM. Fragments of this material outnumber palynomorphs by a ratio of several hundreds or thousands to one. Between three minutes and 36 hours of oxidation with Schulze’s solution (70% HNO₃ supersaturated with potassium chlorate, KClO₃), followed by one subsequent rinse with 2% potassium hydroxide (KOH) solution, were used to liberate palynomorphs from the AOM. The >10 μm kerogen fraction from the interbedded limestones at these levels is composed of palynomorphs, brown and black wood fragments and dark-colour clamped AOM. In general, oxidation of preparations from such limestones was restricted to a nitric wash. In the upper M. nodosoides–C. woollgari Ammonite Zones, the >10 μm kerogen fraction from both mudstones and the relatively argillaceous limestone samples at these levels is dominated by dark-colour, clamped AOM, and extended oxidation with Schulze’s solution, as described above, was applied.
All preparations were stained with Safranin O solution (red stain). Approximately equal portions of organic residues from each sample were strewn over four 22 x 22 mm cover slips, dried and mounted onto microscope slides using Petropoxy 154. ‘Absolute abundance’, i.e. the estimated concentration of palynomorphs in the samples (counts per gram, c.p.g.), was calculated using the volumetric method described by Dodsworth (2000).

Three hundred palynomorphs were counted from one microscope slide for each sample and remaining material from up to four microscope slides was subsequently scanned for additional rare taxa. Relative abundance categories follow Dodsworth (2000, Figures 6, 7, 12) and are described as rare (1, 2 or outside the count), common (3–29) or abundant (30+). Standard palaeoenvironmental parameters were calculated for the Pueblo section, including: (i) the ratio between terrestrial (T) and marine (M) palynomorphs (T:M ratio) as a proxy for terrestrial input; (ii) the ratio between P-cysts and G-cysts (P:G ratio) of the dinoflagellate cyst assemblage as a proxy of nutrient input; (iii) the diversity of the dinoflagellate cyst assemblages was calculated using both Shannon and Simpson indexes. Detailed discussion of these palaeoenvironmental parameters is provided in Eldrett et al. (2017). As mentioned above, the higher 26 samples documented here for the first time received extended laboratory oxidation with Schulze’s solution, which can skew the P:G ratio due to a preferential wholesale removal of G-cysts by relatively severe oxidation (Dodsworth 1995, 2000, 2004a, Figure 4). It should be noted that the prominent modern P-cyst family, Protoperidiniaceae, probably do not extend back stratigraphically before the Campanian Stage of the Cretaceous (Bujak & Davies 1983). Protoperidiniaceae, in contrast to the much more resilient mid-Cretaceous P-cysts recorded here (and other Cenozoic P-cyst families), are readily removed from preparations by relatively mild laboratory oxidation such as a nitric wash (pers. obs.; Harland et al. 1980; Davies et al. 1982) or by natural oxidation of Quaternary sediments below the seafloor (Reichart & Brinkhuis 2003; Zonneveld et al. 2007).

All the Pueblo microscope slides are curated in the collections of the Centre for Palynology, Department of Animal and Plant Sciences, University of Sheffield, UK. A full range chart of palynological data is available in the Supplement. Full author names and synonyms of dinoflagellate cysts can be found in Williams et al. (2017).

3. Results

3.1. Palynology across the Lower–Middle Turonian boundary at Pueblo

A description and discussion of the palynology of the lower 53 samples from Pueblo (Beds 63 to 109) can be found in Dodsworth (2000). In these samples, the purer micritic limestone beds tend to contain lower recovery (200–3000 c.p.g.) uncompressed palynomorphs that are dominated by dinoflagellate cyst assemblages while the mudstone interbeds contain higher concentrations (1000–17,000 c.p.g.) of compressed/flattened palynomorphs (Figure 5). The 26 newly analyzed Pueblo samples (Beds 110 to 125) also yielded predominantly marine phytoplankton assemblages; mainly dinoflagellate cysts together with subordinate prasinophyte phycocysts and foraminiferal test linings. Concentration of palynomorphs is generally higher from Bed 121 (BC65) to Bed 125 (BC80; >10,000 c.p.g. in most mudstone lithologies) than from the Bed 110 (BC54) to Bed 120 (BC64; generally <10,000 c.p.g. in mudstones). Dinoflagellate cyst assemblages are diversified (Shannon–Weiner: >3.0; Simpson–Hunter: >0.8); a total of 153 taxa were recorded from the 26 samples, with between 28 and 75 taxa present in each sample. P-cysts taxa are prominent in most samples (P:G ratio >0.5): *Palaehystrichophora infusorioides* is abundant throughout the section while *Alterbidinium* ‘daveyei’, *Isabelidinium? globosum*, *Isabelidinium magnun*, *Subtilisphaera pontis-mariae* and *Subtilisphaera* spp. are generally common; *Trithyrodinium suspicatum* and *Trityrodinium* spp. are abundant from Bed 121 (BC66) to the top of Bed 125 (BC80); *Eurydinium glomeratum* is common to abundant from Bed 112 (BC57) and above.

G-cysts are mostly represented by gonyaulaccean and areoligeracean taxa. The former include *Spiniferites ramosus*, *Spiniferites twistringiensis*, *Floretiella* spp., *Pterodinium cingu- latum cinguulatum* and sporadic influxes in some limestone beds of *Hystrichodinium pulchrum*, Beds 111, 118 and 120 (BC55, BC62, BC64), and *Tanyosphaeridium* spp., Beds 111 and 113 (BC55, BC58). Areoligeracean taxa include *Tenua hystrix*, abundant from Bed 116 to basal Bed 121 (BC60 to BC65), spanning the Lower–Middle Turonian boundary in Bed 120, *Cauveridinium membraniphorum*, which is common up to Bed 120 (BC64), and *Senoniaaspera turonica*, which is common from Bed 116 (BC60) to the top of the sampled section. Specimens of *Heterosphaeridium conjunctum* and *Heterosphaeridium heteracanthum* are common between Bed 90 (BC34) and Bed 116 (BC60).

Land-derived (terrigenous) pollen and spores are mainly represented by gymnosperm pollen (bisaccate pollen and *Classopolis*), with subordinate angiosperm Normapolles pollen (*Atlantopolis* and *Complexiopollis*) and tricolpate pollen, and pteridophyte/bryophyte spores. Gymnosperm pollen comprise up to a third of total palynological assemblages (T:M ratio >0.25).

3.2. Distribution of reworked palynomorphs at Pueblo

Reworked palynomorphs are rare at Pueblo. Pteridophyte spores reworked from Carboniferous deposits, mainly *Densosorites* spp. and *Lycospora pusilla*, are consistent from Bed 63 (BC02) to Bed 78 (BC18) in the Upper Cenomanian section, and occur sporadically in the Turonian section. Taeniate (striate) bisaccate pollen that are probably reworked from Permo-Triassic deposits are also sporadically recorded.

Isolated specimens of dinoflagellate cysts that are probably reworked from Jurassic deposits were recorded from the Upper Cenomanian section; *Chyroeiophagina chyroeiodes* in Beds 64–65 (BC03), *Scrinidinium* spp. and questionable *Gonyaulacysta jurassica* in Bed 78 (BC13.5), and *Rhynchodiiniopsis cladophora* in Beds 70–71 (BC09) and
Bed 85 (BC26). Isolated specimens of taxa probably reworked from Albian to Lower Cenomanian deposits were recorded from the basalt Middle Turonian section; *Endoceratium turneri* in Bed 121 (BC67 and BC69), *Kiokansium unituberculatum* in Bed 122 (BC70) and *Nyctericytus vitrea* and *Ovoidinium scabrosum/verrucosum* in Bed 122 (BC71).

### 4. Biozonation

Many of the dinoflagellate cyst taxa recorded in the present study have stratigraphical ranges broader than Cenomanian and Turonian. The distribution of well-documented regional/inter-regional range base (First Occurrence, FO) and range top (Last Occurrence, LO) data from this interval is discussed below with reference to Portland-1, Pueblo and the SW Texas sections. Vintage dinoflagellate cyst zonation schemes are based mainly on data from southern England and northern France (Clarke & Verdier 1967; Foucher 1981) and eastern Canada (Williams 1977), but can be broadly applied to the central and southern KWIS. The most recent proposed Central and Northern European scheme (Olde et al. 2015a), which reviews in detail the earlier schemes of the Upper Cenomanian to Lower Coniacian interval, is applied here with modifications, as most marker taxa are present over the studied interval. Zone ages given in the following section are based on an astronomical (obliquity) age model applied to the named bioevents in Portland-1 (Eldrett et al. 2017) and Iona-1 (Eldrett et al. 2015a; Figure 7).

Palynological zonation of non-marine Upper Cretaceous sections in the KWIS (Nichols 1994), including continental settings in Wyoming (Lynds & Slattery 2017) and Utah (Akyuz et al. 2016), are based on pollen. The entire study interval considered here for dinoflagellate cyst zonation (Lower Cenomanian to lowermost Coniacian) is tentatively correlated with the *Nysapollenites albertensis* Singh, 1971 (Pollen) Interval Biozone of Nichols (1994). The presence of marine sections containing dinoflagellate cysts in central Wyoming and southern Montana (Burgess 1971) indicates that the dinoflagellate cyst zonation defined here may be applicable to the coeval Frontier Formation and Greenhorn Formation in those areas.

#### 4.1. Albian–Cenomanian boundary interval

The Dakota Sandstone in Portland-1 contains dinoflagellate cysts from 209.59 m to 205.11 m, including taxa that are indicative of proximity to the Albian–Cenomanian boundary: rare to common *Ovoidinium* spp., mainly specimens transitional between *O. scabrosum* and *O. verrucosum*, and *Senonisphaera microreticulata*, which has a (possibly reworked) LO in the first sample analyzed (204.39 m) above the unconformity at the base of the overlying Graneros Shale. The LO of *O. scabrosum* occurs near the top of the Albian Stage in Northern Europe (Costa & Davey 1992; Duxbury 2001). Brideaux and McIntyre (1975), studying the Northwest Territories in Canada, and Århus (1991), studying north Greenland, southeastern Spitsbergen and the Barents Sea, considered the total stratigraphical range of *S. microreticulata* to correlate with the base and top of the Middle Albian Substage. Williams and Bujak (1985) indicated a broader range for the taxon, coinciding with entire Albian Stage. In Portland-1, the top of the Dakota Sandstone (205.10 m) is assigned a date of ~99 Ma based on carbon isotope correlation with the astronomically calibrated Iona-1 core (Eldrett et al. 2017), which is younger than the estimated age of the base of Early Cenomanian Substage at its GSSP section in France, ca. 100.5 Ma (International Chronostratigraphic Chart, v. 2017/02; http://cretaceous.stratigraphy.org), suggesting that *S. microreticulata* may range into the lowermost Cenomanian in Portland-1.

#### 4.2. The Ovoidinium verrucosum interval zone

**Definition of zone**: From the first sample stratigraphically above the LO of *Senonisphaera microreticulata* (base of zone) to the stratigraphically highest sample below the FO of common *Litosphaeridium siphoniphorum* (top of zone).

**Depth range**: Portland-1, 201.27 m to 178.33 m; Iona-1, base not seen, top at 144.63 m.

**Age of zone**: Early to Middle Cenomanian (ca. 96.8 Ma to 96.4 Ma in Portland-1, Figure 4, Eldrett et al. 2017; top at ca. 96.6 Ma in Iona-1, Figure 7).

**Discussion**: The Graneros Shale in Portland-1, from its base (205.10 m) up to the proposed unconformity at the top of...
the carbon isotope Middle Cenomanian Event (MCE) at ca. 178.20 m (Figure 4; Eldrett et al. 2017), contains dinoflagellate cyst taxa, including rare Ovoidinium verrucosum, with its LO at 178.33 m, and Luxadinium propatulum, LO at 179.86 m. Burgess (1971) reported O. verrucosum to be common over its stratigraphical range in central Wyoming and southern Montana. The LO of O. verrucosum has been used as a Lower Cenomanian marker event in North America and northern Europe (Cookson & Hughes 1964; Burgess 1971; Williams & Bujak 1985; Costa & Davey 1992; Scott et al. 2018, who dated it at 96.43 Ma in Kansas). In Portland-1, the correlation of its LO with the MCE suggests that it ranges into the basal part of the Middle Cenomanian in that section. It is very rare in the Buda Limestone of SW Texas, as are most P-cysts (Cornell 1997; Eldrett et al. 2017; Section 1.2), but is consistently present in the Lower Cenomanian part of the siliciclastic Woodbine Formation in eastern Texas (Beach 1985).

We propose using the stratigraphically highest sample below the FO of common Litosphaeridium siphoniphorum to define the top of the O. verrucosum Interval Zone. In Portland-1, this is coincident with the LO of O. verrucosum (178.33 m), occurring within the MCE. The FO of common L. siphoniphorum in Portland-1 (178.10 m) occurs immediately above the top of the MCE but occurs around the base of the MCE in Iona-1 and Innes-1 (cf. Figures 6 and 7).

4.3. The Litosphaeridium siphoniphorum interval zone

**Definition of zone:** From the FO of common Litosphaeridium siphoniphorum to define the top of the O. verrucosum Interval Zone. In Portland-1, this is coincident with the LO of O. verrucosum, occuring within the MCE. The FO of common L. siphoniphorum in Portland-1 occurs immediately above the top of the MCE but occurs around the base of the MCE in Iona-1 and Innes-1 (cf. Figures 6 and 7).

**Depth range:** Portland-1, 178.10 m to 147.22 m; Iona-1, 144.21 m to 104.83 m.

**Outcrop range at Pueblo:** Bed 63 (BC01; base not seen) to Bed 73 (BC10).

**Age of zone:** Middle to Late Cenomanian (ca. 96.4 Ma to 94.54 Ma). In Portland-1, Figure 4, Eldrett et al. 2017, ca. 96.6 Ma to 94.62 Ma in Iona-1, Figure 7.

**Discussion:** In Central and Northern Europe, Olde et al. (2015a) proposed a Cenomanian Litosphaeridium siphoniphorum Zone, with the top defined by an intra-Upper Cenomanian bioevent, LO common L. siphoniphorum. They did not define a base for the zone. For the central and southern KWIS, we propose using this zone with the additional criterion of using the regional FO of common L. siphoniphorum as its base. The FO of the taxon occurs in Upper Albian deposits in Europe (Davey & Verdier 1971; Foucher 1981) but the FO of consistent L. siphoniphorum is stratigraphically higher than this in the KWIS, occurring in the upper part of the Graneros Shale in Portland-1 and just above the base of the lower Eagle Ford in SW Texas, the FO of common L. siphoniphorum occurring in the same sample or a short distance above (Figure 7). This may also be the case in a proto-Atlantic site at Tarfaya, Morocco (Figure 1), where FO common L. siphoniphorum is calibrated to the upper part of the Rotalipora reicheli Foraminiferal Zone (Prauss 2012a), in proximity to the Lower–Middle Cenomanian boundary.

The LO of consistent L. siphoniphorum is coincident with the LO of common L. siphoniphorum at Pueblo (and with the top of the zone), in Bed 73, but the two events are separated by a short distance in Portland-1 (28 cm; Eldrett et al. 2017), Iona-1 (139 cm; Eldrett et al. 2015a) and Lozier Canyon (70 cm; Dodsworth 2016), possibly reflecting more expanded sections and/or higher-density sampling. In southern England, LO consistent and LO common L. siphoniphorum are coincident in Plenus Marl Bed 6 at Lulworth and Eastbourne (Dodsworth 2000; Pearce et al. 2009). Isolated specimens have sporadically been recorded in the Turonian at Pueblo and elsewhere; the Isle of Wight, southern England (Clarke & Verdier 1967), the Touraine area of France (Foucher 1982) and the Münster Basin of Germany (Marshall & Batten 1988). Foucher (1983) recorded an isolated specimen from Coniacian deposits. It is uncertain whether these isolated Turonian and younger occurrences are as a rare species or as reworking. Previous reports of LO consistent L. siphoniphorum in Lower Turonian deposits (Williams & Bujak 1985; Costa & Davey 1992) derive from former assignment of the Plenus Marl in southern England to this subgroup (Jeffries 1962, 1963); it has subsequently been confidently reassigned to the Upper Cenomanian Substage (Jarvis et al. 1988; Gale et al. 1993). An Upper Cenomanian LO of consistent L. siphoniphorum has been reported worldwide, including other locations in southern England (Davey 1969; Hart et al. 1987), the Witch Ground Graben, central North Sea (Harker et al. 1987), France (Foucher 1979, 1980; Courtinat et al. 1991), northern Spain (Mao & Lamolda 1999), northern Germany (Marshall & Batten 1988), eastern US (Aurisano 1989), Ocean Drilling Project holes including Demerara Rise (Leg 207, Site 1260) and Kerguelen Plateau (Leg 183, Site 1138; Eldrett et al. 2017), Australia (Morgan 1980; McMinn 1988) and New Zealand (Hasegawa et al. 2013; Schieler & Crampton 2014).

**Supplementary bioevents:** Burgess (1971) used the FO of Alterbdinium ‘daveyi’ (i.e. the comparable form Deflandrea sp. cf. D. acuminata) as an intra-Cenomanian correlative event in

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**Plate 2.** Dinoflagellate cysts from Pueblo and Portland-1, Colorado. Specimens were magnified at 400×; the 50 μm scale bar applies to all specimens. Pueblo beds, corresponding BC sample/slide numbers, Portland-1 depths/slides, and specimen England Finder co-ordinates are given. Figure 1. Caueridinium membranophorum (Cookson & Eisenack 1962) Masue in Faunomier & Masue 2004, Bed 110 (BC54A), R30/1; Figure 2. Senonisphaera turonicus (Prössl 1990, ex Prössl 1992) Pearce et al. 2011, Bed 124 (BC76A), S33/3; Figure 3. Senonisphaera rotundata Clarke & Verdier 1967, Bed 113 (BC58A), E38/3; Figure 4. Heterosphaeridium conjunctum Cookson & Eisenack 1968, Bed 112 (BC56A), L31/1; Figure 5. Canninia senonica Clarke & Verdier 1967, Bed 102 (BC44A), V32/0; Figure 6. Subtilisphaera pontis-mariae (Deflandre 1936) Lentin & Williams 1976, Bed 121 (BC67), S32/3; Figure 7. Senonisphaera microreticulata Brideaux & McIntyre 1975, Portland-1, 204.39 m, O15/4; Figure 8. Temnodinium hystrix Eisenack, 1958, Bed 124 (BC76A), Q36/0; Figure 9. Florentinia? torulosa (Davey & Verdier 1976) Lentin & Williams 1981, Bed 102 (BC42A), V52/2; Figure 10. Ovoidinium verrucosum ostium (Davey 1970) Lentin & Williams 1975, Portland-1, 205.15 m, T25/1; Figure 11. Ovoidinium verrucosum verrucosum (Cookson & Hughes 1964) Davey 1970, Portland-1, 205.15 m, X31/4; Figure 12. Subtilisphaera microreticulata Brideaux & McIntyre 1975, Portland-1, 209.59 m, V26/3.
central Wyoming and southern Montana. Singh (1983) suggested a Middle Cenomanian FO for *A. daveyi* and LO for *Kokanisia williamsii* in Alberta, Canada. Williams (2006) cited a date of 96.2 Ma for the LO of *K. williamsii* in the Scotian Basin, Canada. In Portland-1, both events occur at 169.19 m, just above the base of the Upper Cenomanian as picked by Eldrett et al. (2017), ca. 95.6 Ma, following the suggested level for the Middle–Upper Cenomanian boundary (Kennedy & Gale 2006; Ogg et al. 2016; Section 1.1).

Within the *L. siphoniphorum* Interval Zone, there are other correlative bioevents across the central and southern KWIS, including: a lower influx of prasinophyte phycocyanin, an influx of *Bosedinia* cf. sp.1 and sp.3 of Prauss (2012b), an upper influx of prasinophyte phycocyanin, FO *Isabelidinium magnum* and FO *Cauveridinium membraniphorum* (Figure 7). While the first three events appear to be organic rich and facies controlled, probably related to incursion of anoxic–dysoxic proto-Atlantic water-masses from a southerly direction, and are diachronous, i.e. occurring later farther north in Colorado than in SW Texas (Section 1.2), the latter two events are useful supplemental correlative events within the Upper Cenomanian. The FOs of consistent *I. magnum* (154.18 m in Portland-1; Bed 63 [BC01] at Pueblo) and *C. membraniphorum* (147.22 m in Portland-1; Bed 63 [BC02] at Pueblo), taxa more commonly encountered in northerly Boreal water-masses (Eldrett et al. 2014, 2015a, 2017; Van Helmond et al. 2016), are regionally correlative. The FO of *I. magnum* also occurs in the Upper Cenomanian outside the KWIS, eastern Canada, Labrador–Baffin Seaway (Pedersen & Nøhr-Hansen 2014; Nøhr-Hansen et al. 2016) and Northern Europe (Costa & Davey 1992). The FO of *C. membraniphorum* is recorded from deposits as old as Middle Albian (Foucher 1981), but only consistently from the Middle Cenomanian in high-latitude sites (Van Helmond et al. 2016); its consistent FO at higher stratigraphic levels still, i.e. intra–Upper Cenomanian, in many mid-latitude sites such as Pueblo, other sites within the central KWIS (Harris & Tocher 2003), and Texas (Eldrett et al. 2014, 2015a, 2017), may reflect a southerly incursion of a Boreal water-mass to the KWIS and Central Europe, the Plenus Cold Event (Section 1.2).

The FO of *Dinogymnum* spp. occurs at 147.90 m in Portland-1 and Bed 66 (BC04) at Pueblo. Burgess (1971) demonstrated the correlation of its FO (i.e. FO *Dinogymnum heterocostatum*) in proximity to the Frontier Formation–Greenhorn Formation boundary, as defined in central Wyoming and southern Montana.

At Pueblo, the LO of *Pterodinium? pterotum* occurs in Bed 68 (BC06). Davey and Williams (1966) and Clarke and Verdier (1967) recorded it from the Middle and Upper Cenomanian in southern England, its LO coinciding with that of *L. siphoniphorum* within the Plenus Marls on the Isle of Wight. The two events also correlate at Melton Ross in eastern England (unpubl. data, cf. Dodsworth 2000, Figure 12).

### 4.4. The *Cauveridinium membraniphorum* interval zone

**Definition of zone:** From the first sample stratigraphically above the LO of common *Litosphaeridium siphoniphorum* (base of zone) to the higher of either the LO of *Cauveridinium membraniphorum* or the LO of *Isabelidinium magnum* (top of subzone).

**Depth range:** Portland-1, 146.94 m to 145.24 m (shallowest core sample analyzed); Iona-1, 104.18 m to 7.91 m.

**Outcrop range at Pueblo:** Bed 74–75 (BC11) to Bed 125 (BC80; highest sample analyzed).

**Age of zone:** Late Cenomanian to Early Coniacian (ca. 94.62 Ma to 89.45 Ma in Iona-1, Figure 7).

**Discussion:** In Central and Northern Europe, Olde et al. (2015a) proposed a *Cauveridinium membraniphorum* Zone, an interval zone defined between LO common *L. siphoniphorum* (intra–Upper Cenomanian) and LO *C. membraniphorum* (Lower Coniacian). It is composed of a lower unnamed interval, a *Senoniasphaera turonica* Subzone (base marked by the FO of common *S. turonica*) and a *Raetiaedinium truncigerum* Subzone (base marked by the FO of consistent *R. truncigerum*). As the latter taxon is not recorded from Turonian strata in central and southern KWIS sections investigated to date (Figure 7), we propose modifications to the *C. membraniphorum* Zone in this region, splitting it into *Adnatosphaeridium tutulosum*, *Isabelidinium magnum*, *Senoniasphaera turonica* and *Senoniasphaera rotundata* Subzones. As the maker taxon *C. membraniphorum* is rarely recorded above the Turonian Stage in Northern Europe (Costa & Davey 1992) and has not to date been recorded in the KWIS from above the Turonian in Kansas (Scott et al. 2018) or Texas (this paper), we also propose using the alternative LO of *Isabelidinium magnum* to define the top of the zone if *I. magnum* is recorded at a stratigraphically higher level (see Section 4.4.4).

#### 4.4.1. The *Adnatosphaeridium tutulosum* interval subzone

**Definition of subzone:** From the first sample stratigraphically above the LO of common *Litosphaeridium siphoniphorum* (base of subzone) to the LO of consistent *Adnatosphaeridium tutulosum* (top of subzone).

**Depth range:** Portland-1, 146.94 m to 146.18 m; Iona-1, 104.18 m (single sample).

**Outcrop range at Pueblo:** Bed 74–75 (BC11) to top of Bed 78 (BC18).

**Age of subzone:** Late Cenomanian (ca. 94.59 Ma in Iona-1, Figure 7).

**Discussion:** The LO of *Adnatosphaeridium tutulosum* occurs in Bed 78 equivalent at Bunker Hill in Kansas (Harris & Tocher 2003) and in a comparable stratigraphical position at Lozier Canyon outcrop and Innes-1 and ‘Well X’ cores in SW Texas (Figure 7). At locations in Europe, LO *A. tutulosum* also occurs within the uppermost Cenomanian, above LO consistent/common *L. siphoniphorum* in southern England (Plenus Marls, Bed 8 and Bed 6, respectively; Dodsworth 2000; Pearce et al. 2009), eastern England (Dodsworth 1996; 2000, Figure 12), northern Germany (Marshall & Batten 1988), southern France (cf. Courtinat et al. 1991; Jarvis et al. 2011).
and at the same level as LO *L. siphoniphorum* in northern France (Foucher 1983) and Crimea (Dodsworth 2004a).

**Supplementary correlative events include:** LO consistent *Wuroia* spp. occurs at, or just above, the base of the subzone. The taxon occurs between Beds 63 and 73 (BC01 to BC10) at Pueblo and from 147.90 m to 146.94 m in Portland-1. It is also present in the SW Texas sections (Figure 7) but has not been reported outside the KWIS.

LO *Adnatosphaeridium? chonetum* is consistently recorded from Upper Cenomanian strata (Bed 63 to intra-Bed 78; BC01 to BC17) at Pueblo and from 147.90 m to 146.18 m in Portland-1. Harris and Tocher (2003) recorded a similar distribution at Bunker Hill in Kansas, Wahweap Wash in Utah and Blue Point in Arizona. In the SW Texas sections (Figure 7) it also has an uppermost Cenomanian LO, apart from in Iona-1, where its (possibly reworked) LO occurs in basal Turonian strata. The event has not been reported outside the KWIS though the occurrence of *A.? chonetum* in Cenomanian deposits was noted by Davey (1969) in Northern Europe and by Cookson and Eisenack (1962) and Backhouse (2006) in Australia.

A single specimen of *Microdinium setosum* was recorded in Beds 74–75 (BC11) at Pueblo. In Northern Europe, this taxon occurs consistently in the Cenomanian (Clarke & Verdier 1967) but can occur sporadically as high as Middle Turonian in some areas (Costa & Davey 1992).

The FO of abundant *Cyclonephelium compactum* and common *Cauveridinium membraniphorum* occur in Beds 74–75 (BC11) at Pueblo, with an acme of *C. membraniphorum* in Bed 84 (BC25). There is also an isolated influx of common *C. membraniphorum* in the uppermost Cenomanian at Lozier Canyon (32.31 m) and Innes-1 (47.10 m) but at a higher stratigraphical level (78.28 m; Lower Turonian) in Iona-1. In Iona-1, both taxa and forms transitional between them are common in the uppermost Cenomanian to Lower Turonian (72.75 m to 102.95 m; Eldrett et al. 2015a). In Europe, they have a common uppermost Cenomanian to Lower Turonian occurrence in Germany (Marshall & Batten 1988; Prauss 2006), Crimea (Dodsworth 2004a), Poland (Dodsworth 2004b) and France (Foucher 1983; Courtinat et al. 1991). *Cauveridinium membraniphorum* is rare or absent in coeval Tethyan sections farther south, in Spain (Lamolda & Mao 1999; Peyrot 2011; Peyrot et al. 2011, 2012) and Morocco (Prauss 2012a, 2012b).

The LO of *Carpodinium obliquocostatum* occurs in Bed 78 (BC14) at Pueblo. Although rare and sporadic in occurrence, its range top is reported to approximately coincide with that of *A. tutulosum* in Northern Europe (Marshall & Batten 1988; Dodsworth 1996, 2000).

### 4.4.2. The Isabelidinium magnum interval subzone

**Definition of subzone:** From the first sample stratigraphically above the LO of consistent *Adnatosphaeridium tutulosum* (base of subzone) to the first sample stratigraphically below the FO of common *Senoniasphaera turonica* (top of subzone).

**Depth range:** Iona-1, 103.44 m to 59.88 m.

**Outcrop range at Pueblo:** Bed 79 (BC19) to Bed 115 (BC59).

**Age of subzone:** Late Cenomanian to Early Turonian (Colorado) or Middle Turonian (Texas; ca. 94.59 Ma to 91.72 Ma in Iona-1, Figure 7).

**Discussion:** *Isabelidinium magnum* and *Isabelidinium? globosum* are common from Bed 104 (BC46) to the top of the sampled section at Pueblo and there is an influx of prominent *I. magnum* from Bed 106 (BC49) to Bed 108 (BC52). The geographical extent of this influx is not known, as it was not reported in the other central KWIS sections analyzed by Harris and Tocher (2003) and is not present in the SW Texas sections. By contrast, in the northern KWIS, e.g. Saskatchewan and Alberta (Davey 1970; Bloch et al. 1999), both taxa are abundant throughout the Upper Cenomanian to Middle Turonian interval. Along with *Eurydinium glomeratum* (see below), their general southward decline in numbers between Canada, Colorado and Texas probably reflects a cooler more northerly water-mass preference. In Northern Europe, there is a comparable decline in numbers of *I. magnum* between the northerly argillaceous Shetland Group and southerly calcareous Chalk Group (Costa & Davey 1992) while *I.? globosum* and *E. glomeratum* are rare or absent in European sections.

**Supplementary correlative events include:** The LO of *Pterodinium cingulatum reticulatum* occurs in Bed 79 (BC19) at Pueblo. It is common between Beds 64–65 (BC03) and Beds 76–77 (BC12). Clarke and Verdier (1967) also noted a range top around the Cenomanian–Turonian boundary level on the Isle of Wight.

The FOs of *Heterosphaeridium difficile*, *Florentinia buspina* and *Florentinia? torulosa* are Lower Turonian biostratigraphical marker events in Northern Europe (Davey & Verdier 1976; Foucher 1980, 1981, 1983; Tocher & Jarvis 1987; Tocher in Jarvis et al. 1988; Costa & Davey 1992; FitzPatrick 1995; Pearce et al. 2003, 2009). At Pueblo, all three taxa are rare and sporadic in the *I. magnum* Interval Subzone. *Heterosphaeridium difficile* is first recorded in an isolated sample (BC36) from Bed 95 (Lower Turonian) but is more consistent in occurrence in the Middle Turonian, from Bed 122 (BC72). It is a common component of the younger (Coniacian–Santonian) Niobrara Formation in Colorado (unpubl. data). The taxon is first recorded from Middle Turonian deposits in Iona-1 (65.12 m) and Innes-1 (41.78 m). The FO of *F. buspina* at Pueblo has been recorded in Bed 101 by Harris and Tocher (2003) and slightly higher, Bed 105 (BC48), in this study. Harris and Tocher (2003) also recorded its FO in Bed 99 equivalent at Blue Point, Arizona, and Bed 102 equivalent at Bunker Hill, Kansas. It has not been recorded in the Eagle Ford sections in SW Texas. The FO of *F.? torulosa* was used as an intra-Turonian biozonal event by Fitzpatrick (1995). It is sporadically recorded from Bed 102 (BC42) and above at Pueblo, at 24.24 m in Innes-1, and from 24.87 m to 10.75 m in Iona-1.

*Heterosphaeridium* spp., including *H. conjunctum* and *H. heteracanthum*, is common to abundant from Bed 90 (BC34) to Bed 116 (BC60) at Pueblo. The geographical extent of this influx is not known, as it was not reported in the other
central KWIS sections analyzed by Harris and Tocher (2003) and is not present in the SW Texas sections.

Rare and sporadic specimens of questionable *Alterbidinium ioannidesii* occur in Bed 100 (BC40) and above at Pueblo. *Alterbidinium ioannidesii* is a common component of the younger (Coniacian–Santonian) Niobrara Formation in Colorado (unpubl. data). It has not been recorded from the Eagle Ford Group or lower part of the Austin Chalk Formation in the SW Texas sections. Pearce (2010) documented it from the Lower Campanian in Norfolk, eastern England.

The FO of *Canningia senonica* (Plate 2, Figure 5) occurs in Bed 102 (BC43) at Pueblo. Clarke and Verdiér (1967) first published the taxon from Santonian deposits on the Isle of Wight. They mainly differentiated it from the longer-ranging (Upper Cenomanian to Middle Turonian at Pueblo) type species *Canningia reticulata* by a higher reticulum. Santos et al. (2018) considered it to be a junior synonym of *C. reticulata*, based on the co-occurrence of both taxa in Turonian deposits in NE Brazil, and suggested the morphological differences might be linked to palaeoenvironments. However, we retain *C. senonica* because it has a different (shorter) stratigraphical range from *C. reticulata* at Pueblo and in southern England (cf. Clarke & Verider 1967; Pearce et al. 2003, 2009).

At Pueblo, the FO of *Eurydinium glomeratum* occurs in Bed 102 (BC44; intra-Lower Turonian), with the FO of abundant *E. glomeratum* in Bed 112 (BC57). It has a comparable FO at Bunker Hill in Kansas (Bed 98 equivalent; Harris & Tocher 2003). It is common from a stratigraphically lower level, i.e. from the Upper Cenomanian, through to the Middle Turonian in Saskatchewan and Alberta, western Canada (Davey 1970; Bloch et al. 1999). In the southern KWIS, it has a Lower to Middle Turonian range in the relatively expanded, stratigraphically complete section in Iona-1 (98.88 m to 70.45 m; Eldrett et al. 2015a). It was not recorded at Lozier Canyon or Innes-1, where it may be missing in strata eroded in the Lower–Middle Turonian unconformity at the base of the Langtry Member. This correlates with a regional Lower–Middle Turonian hiatus identified throughout much of the KWIS (Ewing 2013) with an estimated duration of >2 million years at Lozier Canyon and Innes-1. The unconformity is much less intense (estimated duration of <0.2 million years) in the more distal Iona-1 core (duration estimates from Eldrett et al. 2015a, 2017).

The FO of *Chatangiella* spp. (rare, sporadic) was recorded from Bed 104 (BC47) at Pueblo in this study. Harris and Tocher (2003) also reported an FO of *Chatangiella spectabilis* from a slightly lower stratigraphic level at Pueblo (Bed 101), Blue Point in Arizona (Bed 98 equivalent) and Bunker Hill in Kansas (Bed 95 equivalent). In SW Texas, it is also first recorded from the Lower Turonian in Iona-1 (74.50 m) and in proximity to the base Middle Turonian sedimentological pick in Innes-1 (43.42 m).

The FO of *Membranilarnacia polycladiata* occurs within the Lower Turonian at Pueblo (Bed 111, BC55) and Iona-1 (79.51 m), and within the Middle Turonian in Innes-1 (34.60 m). However, it was originally published from Albion deposits in Australia (Cookson & Eisenack 1958; Eisenack 1963) and has been recorded throughout the Cenomanian and Turonian in NW Germany (Prössl 1990).

### 4.4.3. The Senoniasphaera turonica interval subzone

**Definition of subzone:** From the FO of common *Senoniasphaera turonica* (base of subzone) to the first sample stratigraphically below the FO of common *Senoniasphaera rotundata* (top of subzone).

**Depth range:** Iona-1, 58.74 m to 40.35 m.

**Outcrop range at Pueblo:** Bed 116 (BC60) to Bed 125 (BC80; the highest sample analyzed).

**Age of subzone:** Early Turonian (Colorado) or Middle Turonian (Texas) to Late Turonian (ca. 91.72 Ma to 90.17 Ma in Iona-1, Figure 7).

**Discussion:** Pearce et al. (2003, 2011) differentiated *Senoniasphaera turonica* from *Senoniasphaera rotundata* by the former’s possession of a relatively thickened ectophragm and greater ornament development on the ventral surface. Specimens close to the holotype of *S. rotundata* (Clarke & Verdiér 1967, Plate 14, Figure 2) have relatively continuous reticulum ectophragm development across the dorsal surface. Both species exhibit a wide range of morphological variation (Prince et al. 1999; Pearce et al. 2011), particularly in regard to extent of the ectophragm reticulum coverage. Both species are present in the central and southern KWIS. Many KWIS specimens have been subjected to ectophragm bleaching caused by the laboratory oxidation applied to most samples, which in some cases hinders consistent and confident differentiation between the two species.

In Europe, *S. turonica* is occasionally recorded from Middle and Upper Cenomanian deposits (Pearce et al. 2011) but it shows a widespread FO consistent and FO common occurrence in intra-Lower or basal Middle Turonian deposits (Foucher 1980; Tocher & Jarvis 1987; Jarvis et al. 1988; Lamolda & Mao 1999; Pearce et al. 2003, 2009). Within the Turonian at Banterwick Barn, Berkshire, southern England, two main occurrences of *S. turonica* were identified by Pearce et al. (2003, 2011): An intra-Lower to Middle Turonian interval, separated from an intra-Middle to Upper Turonian interval by a section in which the taxon is not recorded. The intra-Lower to Middle Turonian occurrence, between the ‘Lulworth’ and ‘Round Down’ marl beds, which can be correlated with Portland-1 and Iona-1 on carbon isotope criteria (Figure 6), shows a comparable distribution to that at Pueblo where FO consistent (Bed 113, BC58) and FO common *S. turonica* (Bed 116, BC60) occur above the OAE-2 isotope event, within the upper part of the Lower Turonian. In Texas, FO consistent and FO common *S. turonica* occur at a higher stratigraphical level, at the ‘Pewsey’ carbon isotope event, as calibrated in Iona-1 (Eldrett et al. 2015a) and Innes-1 cores (cf. Figures 6 and 7), possibly correlating with the base of the intra-Middle to Upper Turonian occurrence at Banterwick Barn.

We follow the definition of Olde et al. (2015a) for the base of the *S. turonica* Interval Subzone as the FO of common *S. turonica* but acknowledge that it is picked at different...
stratigraphical levels in Colorado and Texas, as discussed above. For central and southern KWIS, we differ from Olde et al. (2015a) in placing the top of the subzone at the FO of common *S. rotundata*.

**Supplementary correlative events include**: The LO of common *Cauveridinium membraniphorum* occurs in Bed 120 (BC64) at Pueblo and 72.77 m in Iona-1. It provides a useful regional proxy for the Lower–Middle Turonian boundary.

   The FO of abundant *Trithyrodinium susppectrum*/sp. occurs in Bed 121 (BC66) at Pueblo, an event of potential correlative value within the central KWIS. However, in SW Texas, the genus *Trithyrodinium* is common to abundant at several levels throughout the Middle Cenomanian to Upper Turonian section (Eldrett et al. 2015a; Dodsworth 2016).

   At Pueblo, a single specimen of questionable *Raphidodinium fucatum* was recorded in Bed 124 (BC74), a dinoflagellate cyst taxon with a Middle Turonian range base in Northern Europe (Costa & Davey 1992).

4.4.4. The *Senonia* sp. *rotundata* interval subzone

**Definition of subzone**: From the FO of common *Senonia* sp. *rotundata* (base of subzone) to the higher of either the LO of *Cauveridinium membraniphorum* or the LO of *Isabelidinium magnum* (top of subzone).

**Depth range**: Iona-1, 37.89 m to 7.91 m (the shallowest rich recovery core sample analyzed).

**Age of subzone**: Late Turonian to Early Coniacian (ca. 90.17 Ma to 89.45 Ma in Iona-1, Figure 7).

**Discussion**: Although rare specimens of *Senonia* sp. *rotundata* are recorded in the Bridge Creek Member at Pueblo (from Bed 113, BC58), and the upper Eagle Ford Group in Iona-1 and Innes-1, they are relatively consistent and common in the Austin Chalk in the latter localities (Figure 7). In Colorado, they are common in the (Coniacian–Santonian) Niobrara Formation (unpubl. data). At Lozier Canyon, a single sample with common *S. rotundata* occurs at a stratigraphically lower level than in Iona-1 and Innes-1, within the Langtry Member of the upper Eagle Ford Group (Dodsworth 2016) but still within the Upper Turonian. The KWIS observations are in agreement with the those of Pearce et al. (2003, 2011), who noted that specimens close to the holotype of *S. rotundata* are more typical of Upper Turonian and Coniacian deposits at Banterwick Barn in England, although the genus is mainly represented by *S. turonica* in Lower and Middle Turonian deposits there. Pearce et al. (2003) reported an overlap within Upper Turonian deposits of the common occurrences of *S. rotundata* and *S. turonica*. Notwithstanding the issues of consistent identification and potentially diachronous basal common occurrence within the Upper Turonian in the KWIS, we propose an *S. rotundata* Subzone, defined by the FO of its common occurrence.

   The LO of *Cauveridinium membraniphorum* occurs in the Lower Coniacian in Europe (see Olde et al. 2015a for a review), though it is relatively rare and sporadic above the top of the Turonian Stage, which has given rise to its LO being alternatively reported as top Turonian by some European workers (Costa & Davey 1992). *Cauveridinium membraniphorum* is present in some Austin Chalk core samples analyzed from Iona-1 (LO at 10.75 m) and Innes-1 (LO at 25.98 m). The alternative bioevent for the top of the zone/subzone is LO *Isabelidinium magnum*, which also has a Lower Coniacian LO in Northern Europe (Costa & Davey 1992) and is recorded in shallower rich recovery core samples in Iona-1 (7.91 m) and Innes-1 (17.45 m) and the highest outcrop sample at Lozier Canyon (57.15 m). This supports an age not younger than Early Coniacian for the sampled lower part of the Austin Chalk, in agreement with macrofossil evidence for an Austin Chalk age not younger than Coniacian (Cobban et al. 2008) and the astronomical age model for the Austin Chalk from Iona-1 that places the base Coniacian at ca. 21 m depth; with the FO of the calcareous nannofossil *Marthasterites furcatus* Bukry, 1969 at 18.67 m (Eldrett et al. 2015a). Note that the subsequent assignment of the ‘Navigation’ carbon isotope event to the negative δ¹³C excursion overlying the Hitchwood/Late Turonian-Coniacian events in Iona-1 (ca. 44 m; Eldrett et al. 2017) is not consistent with this base Coniacian age assignment and therefore may more accurately represent a negative isotopic shift within the Late Turonian Hitchwood event (Figure 6).

5. Discussion

5.1. The Lower–Middle Turonian substage boundary at Pueblo

Although a regional unconformity is developed around the Lower–Middle Turonian boundary interval over much of the southern and central KWIS (see Section 4.4.2), Kennedy et al. (2000) noted that there is no macrofossil or sedimentary evidence in the pelagic, climatically driven, rhythmically bedded sequence at Pueblo for stratigraphic gaps or for condensation in proximity to the proposed substage boundary level, Bed 120. This is also the case with palynology, as there are no sudden changes in assemblage composition around Bed 120. However, there is an increase in areoligeracean dinoflagellate cyst taxa, mainly abundant *Tenua hystrix*, from Bed 116 to basal Bed 121 (Section 3.1), spanning the proposed boundary. Areoligeraceae may have a relatively nearshore affinity (Brinkhuis & Zachariasse 1988; Harker et al. 1990, p. 202–204; Li & Habib 1996; Olde et al. 2015b) and their influx could reflect an interval of relatively low sea level. This interpretation is supported by a minor influx of reworked dinoflagellate cysts probably derived from Albian to Lower Cenomanian deposits in Beds 121 and 122 (Section 3.2). Higher palynomorph concentration in Beds 121 to 125 (Section 3.1), including >1000 gymnospore pollen c.p.g. in many samples (Figure 5; Supplement), could also be considered consistent with greater dilution of biogenic carbonate by an increased supply of terrigenous mud during deposition of the relatively argillaceous Middle Turonian part of the Bridge Creek Member. A broad increase in prasinophyte relative and absolute abundance from Bed 98 to Bed 125 (Figure 5; Supplement) may reflect sea-level highstand and subsequent fall after maximum Greenhorn transgression (cf. Prauss 1993, 2000).
5.2. Comparison with the Western Canada sedimentary basin

The upper part of the Belle Fourche Formation and the Second White Specks Formation in Saskatchewan and Alberta, western Canada, contain palynological assemblages that are comparable to those of intra-Early to Middle Turonian age in this study (Section 4.4.2), including abundant *Isabelidinium magnum, Isabelidinium? globosum* and *Eurydinium glomeratum*, and *Heterosphaeridium difficile*, which is rare in the upper Belle Fourche but can become abundant in the Second White Specks (Bloch et al. 1999). These deposits were tentatively dated as Late Cenomanian to early Middle Turonian in age by Bloch et al. (1999) but our calibration to formal substages suggests that their age range may be younger, i.e. intra-Early to Middle Turonian. Alternatively, common occurrences of these Boreal taxa at Pueblo may reflect much later migration into the central KWIS than their earliest common occurrences in western Canada, as suggested in Section 4.4.2.

6. Conclusions

1. A dinoflagellate cyst zonation is proposed for the Western Interior of the USA, comprising three zones and four subzones. In the Iona-1 (SW Texas) and Portland-1 (Colorado) cores, the zonation is calibrated by radiometric dates from bentonites and astronomical age modeling of the inter-bedded sedimentary strata. Available published biostatigraphical data from molluscs, foraminifera and calcareous nannofossils, and inter-regional carbon isotope stratigraphy, further constrain the zonation at the Pueblo GSSP outcrop (Colorado) and three other sections in SW Texas.

2. Although the zonation provides a working scheme that is potentially applicable to marine sections throughout the Western Interior, regional differences in several taxa are recognized. Some diachronous dinoflagellate cyst events probably followed incursion of proto-Atlantic Tethyan Realm water-masses from south to north; onset *Bosedinia* abundance during the Middle Cenomanian in Texas pre-dates that in Colorado (Late Cenomanian). Conversely, some events probably followed boreal water-masses southwards; onset *Isabelidinium/Eurydinium* abundance in Canada (Late Cenomanian?) may pre-date that in Colorado (Early Turonian).

3. The base of the *Senoniasphaera turonica* Interval Subzone cannot be treated as an isochronous level across the central and southern KWIS. It occurs in the upper part of the Lower Turonian at the Pueblo GSSP but is stratigraphically higher, in the upper part of the Middle Turonian, in the SW Texas cores Iona-1 and Innes-1. It is possible that *S. turonica* migration into the region occurred in two main pulses (as seen in Banterwick Barn, England), only the later event reaching Texas.

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Disclosure statement

No potential conflict of interest was reported by the authors.

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