

**STRATIGRAPHY AND DEPOSITIONAL SETTING OF THE WITPOORT
FORMATION (WITTEBERG GROUP, LATE DEVONIAN) IN THE
EASTERN CAPE, SOUTH AFRICA**

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A thesis submitted to the Faculty of Science, University of the Witwatersrand, Johannesburg, in
fulfilment of the requirements for the degree of Doctor of Philosophy

DECLARATION:

I declare that this thesis is my own, unaided work*. It is being submitted for the Degree of Doctor of Philosophy at the University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination at any other University.

A handwritten signature in black ink, appearing to be 'M. M. M.', written on a horizontal line.

(Signature of candidate)

On this 30th day of May 2025

*except where stated in text

No problem of geology compares in importance with the question of the origin of bedding

Walther (1894)

translated from original German in Campbell (1965, p. 7)

ABSTRACT

The Witpoort Formation, a Late Devonian (Famennian) coastal siliciclastic sedimentary succession, has not received a detailed sedimentary facies analysis in the past three decades, despite significant palaeontological discoveries from its strata. Here, sedimentary facies, sediment dispersal patterns and ichnology are, for the first time, applied in combination to form a sedimentological background to the only high-latitude palaeoecological dataset from its time stage. The results of this thesis indicate a non-marine, sheet-braided fluvial deposit in which transgression of the coast led to the development of marginal-marine, estuarine facies, favouring the preservation of body and trace fossils. The deposits accumulated in erosively-based, valley-fill fluvio-estuarine cyclothems along an approximately WNW–ESE trending shoreline bordering the Agulhas Sea.

Towards the top of the succession (Perdepoort Member) fluvial progradation, and the lack of marginal-marine facies, indicate maximum regression, consistent with a globally recognised sea-level fall during the late Famennian. The end-Devonian Hangenberg Event is contemporary with the uppermost strata of the Witpoort Formation, in concord with prior assessments. Evidence for glaciation associated with this event in South Africa remains tenuous, though it is predicted that careful study of the Skitterykloof Member in the uppermost Witpoort Formation will find that periglacial processes were operating during its deposition, and perhaps also lower in the stratigraphic column. The Kellwasser Event, defining the Frasnian–Famennian boundary, is not recognised in the studied stratigraphic column, though it is predicted that the event will have occurred during deposition of strata that underlie the Witpoort Formation.

Analysis of eustatic patterns from the global Famennian stratigraphic record and comparison with the stratigraphy described herein indicates that eustasy formed an important control on sedimentation in the Witpoort Formation. This presents the strongest evidence for

constraining the chronology of its strata. There is an urgent need for further multi-proxy analyses, especially geochemical and palynological, as have been conducted in Euramerican strata, to increase the chronostratigraphic resolution and recognition of Late Devonian extinction events in the Witteberg Group.

Fossil biodiversity of the Witpoort Formation is principally known from the Waterloo Farm locality, which is here shown to be lower in the stratigraphic column, and therefore older, than previously thought. Waterloo Farm is here positioned in the upper Rooirand Member. This deposit appears to have formed in a fresh- to brackish-water inner estuary, possibly tens of kilometres from the shoreline. Analysis of a new fossil locality, Rabbit Ridge, from lower in the stratigraphy, presents a comparable estuarine depositional environment to Waterloo Farm. Higher degrees of marine influence at Rabbit Ridge are indicated by the invertebrate and trace fossil signatures. The ecology of invertebrate communities and trace fossils indicate variable salinity levels and depositional rates from a spectrum of estuarine sub-environments that occurred at intervals throughout the Rooirand Member.

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CHAPTER 1 – GENERAL INTRODUCTION

Since Plumstead 1967 opened her monographic work on the Devonian plants of South Africa with the statement that it was “an unenviable task reviewing the work of scientific pioneers”, it was with historical irony that her work would later be considered pioneering. This third decade of the Twenty-First Century is an opportune moment to present a new investigation into Devonian rocks and fossils in South Africa, and specifically those of the Late Devonian, which Plumstead and contemporary geologists were only beginning to understand. The South African Late Devonian strata belonging to the Witteberg Group, upper unit of the tripartite Cape Supergroup (Fig. 1.1), forms one of the less-studied stratigraphic units from the country, and one for which, in Plumstead’s day, there was little impetus to understand. What follows in this thesis is the first concerted investigation into strata of the South African Late Devonian this millennium, one that is urgently required, especially in light of new, and forthcoming, palaeontological evidence from the Witpoort Formation. At the beginning of this study the Witpoort Formation was envisaged as a shallow-marine deposit characterised by stacks of tabular, supermature quartzitic sandstone deposited during the Famennian Stage. The sandstone strata are held as shoreface, beach and tidal channel deposits from a storm-dominated Gondwanan shoreline, but their diagnostic sedimentary structures are only vaguely understood, and it could be contended that the sequence stratigraphy of its lithologically monotonous sandstones is virtually unknown (Hiller and Taylor 1992; Thamm and Johnson 2006; Harris *et al.* 2021a).

Devising the ‘origins of bedding’ is a discipline that has made great strides since 1894 when Walther proposed the Law of Succession of Facies. Progress in great leaps with understanding sedimentary facies successions and the time scales of their deposition continues, for example with actuo-sedimentology of tidalites in the 1960s and ‘70s by German and Dutch schools. With the identification of neap-spring tidal bundles by Visser

(1980) the keys to diagnosing tidal processes unlocked a bloom in recognising ancient tidalites (Williams 1989; Archer *et al.* 1995). Then out of the same school came a breakthrough in recognising the deposits of interacting river floods and the estuarine turbidity maximum in a fluvial-tidal transition zone (Van den Berg *et al.* 2007), which led a surge of studies recognising the importance of this zone in ancient sedimentary rocks (e.g. Dalrymple and Choi 2007; Martinius and Gowland 2011; Dalrymple *et al.* 2015). In parallel to this the biological responses to salinity changes as reflected in trace fossils are a highly sensitive proxy for salinity changes (Buatois *et al.* 1997b, 2005; Prescott *et al.* 2014; La Croix and Dashtgard 2015), perhaps the most reliable salinometer in the sedimentologist's toolkit (Dalrymple and Choi 2007). Though fifty years ago the “reliable determination of the tidal sub-environment ... seems to be virtually impossible for ancient clastic deposits, except for a few cases.” (de Raaf and Boersma 1971, p. 479), with detailed palaeocurrent analysis, ichnology and the increasing recognition of diagnostic sedimentary structures, extrapolation of tidal processes in ancient deposits has become more plausible. This is but an example of the progress of one sub-discipline rooted in a vast field of sedimentological research.

The Devonian Period is popularly known as the ‘Age of Fishes’, but it might equally well be considered the age in which plants invaded the land (Gensel and Edwards 2001). Plant diversification during the Devonian fundamentally altered landscapes (Davies and Gibling 2010), increased rates of carbon burial, decarbonising and oxygenating the atmosphere (Algeo and Scheckler 1998; Berner 2004; Beerling and Berner 2005). Forestation of the globe and its consequences for the carbon cycle, beginning in the Middle to Late Devonian is considered to have altered the climate, resulting in a three-quarters decrease of atmospheric CO₂ levels by the Early Carboniferous (Beerling and Berner 2005; Franks *et al.* 2014), which has since remained relatively constant due to mediation by forests (Franks *et al.* 2014). This

plant excursion set the stage for a Late Devonian radiation of aquatic tetrapod fishes and the subsequent rise of terrestrial vertebrates (Clack 2012).

The Late Devonian Epoch was a time of climate change and pulsatory extinction events, for which the causes have been debated for decades, and continue to be (Copper 1986; Algeo and Scheckler 1998; Caplan and Bustin 1999; Bond and Wignall 2008; Becker *et al.* 2016; Bábek *et al.* 2016; Carmichael *et al.* 2019; Fields *et al.* 2020; Lakin and Marshall 2020; Rakocinski *et al.* 2021; Smart *et al.* 2023). Understanding the global expression of these events is inhibited by a bias towards low to mid-palaeolatitude systems found in present day North America, Europe, Asia, and Australia. High-palaeolatitude deposits of that age are exclusively found in western Gondwana, which are exposed in present day South America, the Falkland Islands, South Africa and West-Central Africa (Issacson *et al.* 2008). These regions have received significantly less attention in Late Devonian studies than those of Laurussia (Carmichael *et al.* 2019). Furthermore, Late Devonian strata in South America, the Falkland Islands and in northernmost outcrops of the Cape Fold Belt appear to have been truncated by subglacial erosion during the latest Devonian and/or Permo-Carboniferous Periods (Marshall 2016; de Vargas *et al.* 2020; Lakin *et al.* 2021). The Witpoort Formation is, perhaps uniquely, a continuous sub-polar siliciclastic sedimentary record spanning the Famennian Stage (372.2–358.9 Ma; Cohen *et al.* 2013), uppermost of the Devonian Period (Streel and Theron 1999; Thamm and Johnson 2006).

Decades of palaeontological research on the Waterloo Farm fossil lagerstätte, principally by Dr Robert W. Gess, and his discovery of other fossil localities in this formation now provides fresh impetus to understand the genesis of the fossils and their host rocks. Fossils from the Witpoort Formation provide a unique example of Famennian sub-polar biodiversity, which is critical to understanding the effects of latitude on biodiversity, extinction and recovery in the context of the Late Devonian biotic crisis. The formation preserves possible

high-latitude originator taxa such as the oldest examples of naiaditid bivalves (Scholze and Gess 2017) and falcetid sharks (Gess 2024), relict taxa including antarctilamnid sharks (Gess and Coates 2015b), herbaceous plants typical of Middle Devonian Laurussian ecosystems (Harris *et al.* 2021b), and the youngest record of lingulid brachiopods of the genus *Dignomia* (this work), as well as cosmopolitan Famennian plants and vertebrates (Gess and Whitfield 2020). One black shale bed from Waterloo Farm hosts one of the most diverse vertebrate and plant fossil assemblages of its age, including the only sub-polar Devonian tetrapod remains (Gess and Ahlberg 2018).

The Witpoort Formation (Witteberg Group, Cape Supergroup) is a metasedimentary unit that is exposed in a thin band along the northern flank of the Cape Fold Belt, extending for approximately 800 km and estimated (prior to this study) to be up to 850 m thick (Johnson *et al.* 2006) in the Eastern Cape. With few exceptions in the past 85 years (e.g. Mountain 1964) primary sedimentological data of the Witpoort Formation is contained in postgraduate dissertations (Theron 1961; Loock 1967; Johnson 1976; Wright 1981; Malgas 1989; Taylor 1990; Gess 2003; Olivier 2010; Vorster 2013; Harris 2017; Chabangu 2023) and resulting publications (Johnson 1991; Hiller and Taylor 1992; Gess and Hiller 1995a; Olivier *et al.* 2011; Harris *et al.* 2021a) and to a degree in explanations to geological map sheets by the Geological Survey of South Africa (e.g. Theron *et al.* 1991; Council for Geoscience 1995). Since Hiller and Taylor (1992) provided evidence for back-barrier lagoonal shales interbedded in the stratigraphic package, there has been a clear consensus that the Witpoort Formation was deposited in shallow to marginal-marine settings (Shone and Booth 2005; Thamm and Johnson 2006; Gess and Whitfield 2020), with the sole exception of Broquet (1992), who, in his review of the Cape Supergroup raised concerns about the shortage of evidence for marine deposition in the Witpoort Formation and stratigraphically higher and lower orthoquartzite units. Cooper's (1986) event stratigraphy model correlated the Witpoort

Formation to a Famennian sea-level fall based on the Devonian sea-level curve of Johnson *et al.* (1985). Accordingly, subsequent research places the Devonian–Carboniferous boundary at the top of the Witpoort Formation (Theron 1994; Gess 2016).

Based on the above, it is hypothesised that eustasy, as opposed to variation in rate of subsidence, was the dominant control on shallow-marine sedimentation in the Witpoort Formation, predicting that sea-level fluctuations of the Late Devonian and Early Carboniferous Periods will be reflected in the stratigraphic package, which potentially includes the short-term, climate-driven eustatic perturbations associated with extinction events (Bond and Wignall 2008; De Vleeschouwer *et al.* 2013; Kaiser *et al.* 2016; Wichern *et al.* 2024). The aims of the present work are therefore, 1) to assess the palaeoenvironments of the Witpoort Formation, 2) to test, and potentially refine, Cooper’s event stratigraphy method for dating the Witpoort Formation with reference to subsequent work on Late Devonian eustasy and 3) to revise the stratigraphy of the fossil localities of the Witpoort Formation.

The following chapter reviews key literature on the geological setting of the Witpoort Formation and its Late Devonian context, which will provide the background to the results from stratigraphic logging (Chapter 4), sedimentary facies analysis (Chapter 5), ichnology (Chapter 6) and a new fossil locality from the formation that is described in Chapter 7. The results of the thesis and their implications for Devonian studies will be discussed in Chapter 8.

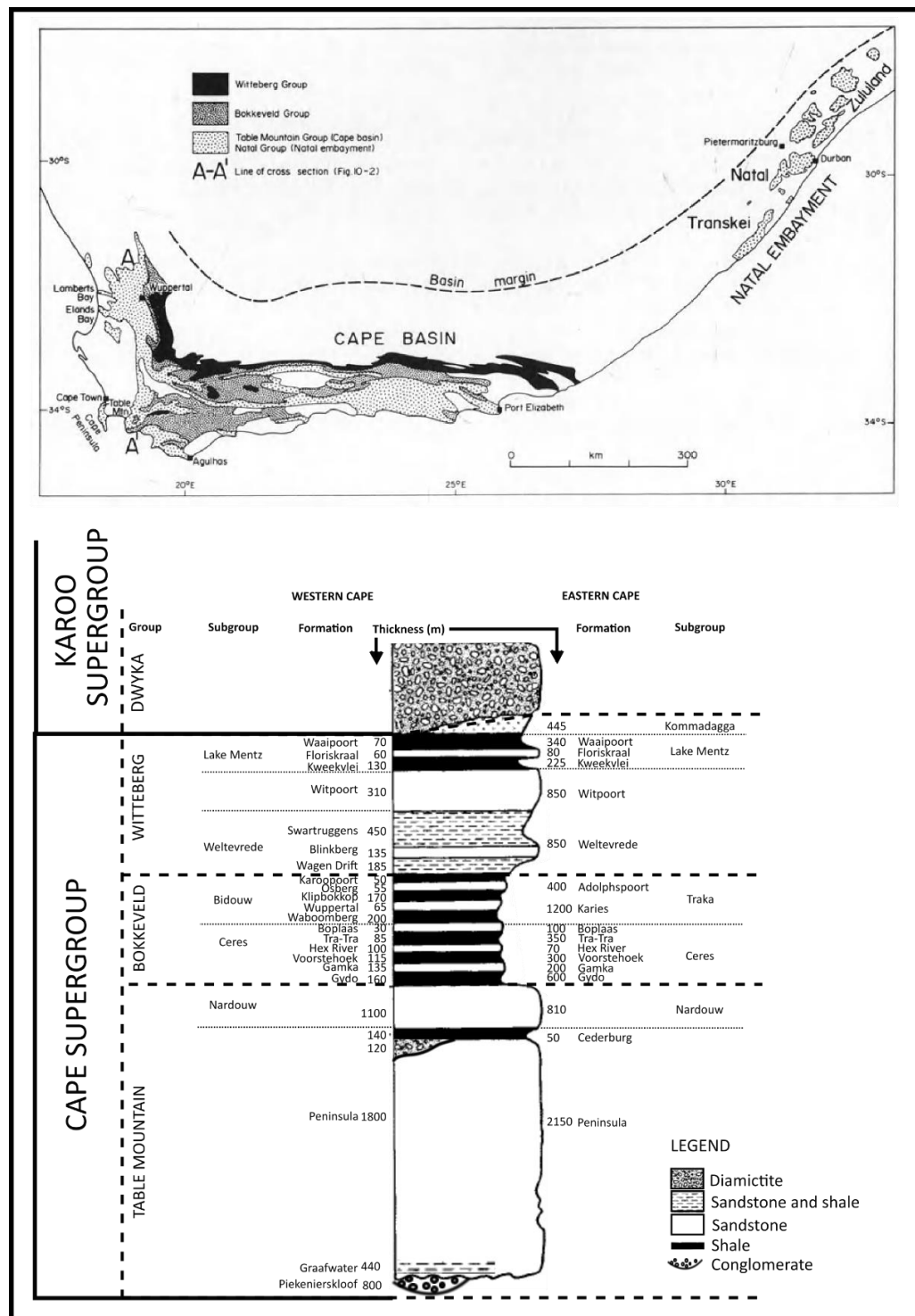


Figure 1.1: Map and stratigraphic column for the Cape Supergroup. Map (above) reproduced from Tankard *et al.* (1982). Stratigraphic column (below) modified from Theron (1994) with thicknesses of stratigraphic units as given by Tankard *et al.* (2012).

CHAPTER 2 – LITERATURE REVIEW

2.1. The Late Devonian Epoch: climate change, extinction and sea level

The Late Devonian includes the Frasnian Stage (382.7–372.2 Ma) and the Famennian Stage (372.2–358.9 Ma) (Cohen *et al.* 2013). Greenhouse conditions of the Silurian and Devonian Periods gave way to the Late Palaeozoic Ice Age with a latest Devonian global cooling episode during the Famennian Stage (Joachimski *et al.* 2009), concurrent with falling sea levels. Analyses by various proxies project a eustatic fall beginning around the Frasnian–Famennian boundary following peak sea levels in the Middle to early Late Devonian (Johnson *et al.* 1985, 1986; Sandberg *et al.* 2002; Haq and Schutter 2008; Becker *et al.* 2012; Marcilly *et al.* 2022; Van der Meer *et al.* 2022). Based on estimates of flooded land area through the Phanerozoic, Marcilly *et al.* (2022) projected a sea-level drop of 124 m during the Famennian between 370 and 360 Ma. A more modest sea-level fall of less than 30 m, for the same time interval, is projected by Van der Meer *et al.* (2022) based on tectonic modelling (Sr isotopes) and glacioeustasy. Based on a lithological section in the Catskills, Haq and Schutter (2008) recognise eight third-order eustatic cycles within the context of a long term, second order (>5 Myr) sea-level fall during the Famennian. Similarly, Becker *et al.* (2012) projected eight third-order cycles on top of an overall shallowing ‘megacycle’ that began in the latest Frasnian. Shorter term, third-order (0.5–5 Myr duration) eustatic signals exhibit more variation because they are prone to the effects of syn-sedimentary tectonic processes and the diverse interpretations of geologists. The literature on Famennian third-order sea-level change from a wide range of stratigraphic sections is reviewed in Chapter 8.5.

Famennian climate change culminated in glaciation at the end of the Devonian Period, producing the first glacial deposits known since the Hirnantian (Late Ordovician), 80 million years prior (Copper and Scotese 2003). Latest Famennian glacial deposits are recognised from Gondwanan sub-polar strata in present-day Bolivia, Brazil, West-Central Africa and

possibly South Africa (Caputo 1985; Isaacson *et al.*, 2008; Almond *et al.* 2002; Lakin *et al.* 2021), extending into subtropical palaeolatitudes in the Appalachian Basin (Brezinski *et al.* 2010; Ettensohn *et al.* 2023). A contemporaneous glacio-eustatic regression is evidenced by a third-order sequence boundary marked by the Hangenberg Sandstone (Rhenish Massif) and its equivalents (Kaiser *et al.* 2016; Denayer *et al.* 2021), and reports of incised valleys across North America and Europe (van Steenwinkel 1993; Brezinski *et al.* 2010). These corresponding latest Famennian (Strunian) regressive event beds are correlated at a global scale by miospore and conodont biostratigraphy (Streel *et al.* 2013; Becker *et al.* 2016; Lakin *et al.* 2021) and are suggested to indicate a glacio-eustatic sea-level fall of approximately 100 m at the end of the Devonian Period (Haq and Schutter 2008; Kaiser *et al.* 2011; Kaiser *et al.* 2016), although the scale of sea-level fall is not universally agreed upon (e.g. Denayer *et al.* 2021). Evidence for glaciation predating the latest Famennian was proposed by Isaacson *et al.* (2008), indicating glacial conglomerates as old as Middle Famennian in Bolivia and Brazil. Lakin *et al.* (2016, 2021) argued against this claim, finding no biostratigraphic evidence for Famennian glaciation in South America predating the latest Famennian (but see Wicander *et al.* 2011). Some research has hypothesised glacio-eustatic sea-level fluctuations beginning in the Late Frasnian Stage, but there remains no evidence for glacial deposits during this interval (Sandberg *et al.* 2002; McClung *et al.* 2013; McGhee 2014).

The final stages of the Devonian were characterised by climatic instability with periodic ocean anoxic events, related to climatic and eustatic perturbations that affected marine and terrestrial ecosystems (Bond and Wignall 2008; Kaiser *et al.* 2016). The most impactful of these anoxic events were the Kellwasser Event near the Frasnian–Famennian boundary and the Hangenberg Event at the end of the Famennian. Researchers continue to search for causes for the Late Devonian extinction events (e.g. Bond and Wignall 2008; Becker *et al.* 2016; Kaiser *et al.* 2016; Percival *et al.* 2019; Carmichael *et al.* 2019; Wichern *et al.* 2024).

2.1.1. The Kellwasser Event

The Kellwasser Event in the latest Frasnian is widely referred to as the second of the of the ‘Big Five’ extinction events (e.g. Bond and Wignall 2008; Ma *et al.* 2022). The Big Five concept was based on a broad review by Raup and Sepkoski (1982), which may have lumped the Late Devonian extinctions together (Carmichael *et al.* 2019). Apparent loss of vertebrates during this event may be related to a ‘backstripping’ effect from the terminal Devonian Hangenberg Event (Sallan and Coates 2010). Kellwasser has been referred to as a mass depletion rather than an extinction event, related to the proliferation of invasive species and a reduction in species origination (Stigall 2012; Stanley 2016). The event resulted in the widespread demise of Frasnian stromatoporoid reef communities and warm water adapted marine invertebrate taxa (Copper 1986; Copper and Scotese 2003; Bond *et al.* 2013).

The event is named after a lithological section at Kellwasser Kalke within the Harz Mountains in Germany (Carmichael *et al.* 2019). The event is marked by a pair of black shale beds at the type section, the Lower and Upper Kellwasser horizons, which are correlated across several basins worldwide in the *rhenana* and *linguiformis* conodont zones respectively (Bond and Wignall 2008; Carmichael *et al.* 2019). Positive excursions in $\delta^{13}\text{C}$ associated with these horizons have been recognised at a global scale (Joachimski *et al.* 2002). Correlatable black shales are widespread within deep-marine sections, mostly within low to mid-palaeolatitude basins although records from basins outside of this geographic zone, as well as its recognition in shallow water settings (Bond *et al.* 2013; Mottequin and Poty 2016), confirm the global scope of the event (Carmichael *et al.* 2019).

Recognition of the Frasnian–Famennian (F–F) boundary based on conodont biostratigraphy (*linguiformis*–*triangularis* boundary) places it at the top of the Upper Kellwasser Horizon (Bond and Wignall 2008). This boundary is dated to 371.93–371.78 Ma based on high precision radiometric dating (Percival *et al.* 2018). The Lower Kellwasser and

Upper Kellwasser Events are separated by approximately 600 kyr (De Vleeschouwer *et al.* 2017) or 450 kyr (Pippenger *et al.* 2023) and lasted approximately 250 and 100 kyr respectively (Wichern *et al.* 2024). The timing of these events has been related to orbital forcing and related climate change during an interval of approximately a million years (De Vleeschouwer *et al.* 2017; Ma *et al.* 2022; Wichern *et al.* 2024).

Marine anoxia is widely held as the direct cause of F–F extinctions (Smart *et al.* 2023), but the underlying driving mechanism is not, as yet, clearly understood. Late Devonian volcanism has gained recent attention due to mercury enrichments recorded in the Upper Kellwasser Horizon in many sections (e.g. Racki *et al.* 2019; Ernst *et al.* 2020; Liu *et al.* 2021; Zhao *et al.* 2022; Smart *et al.* 2023; Pippenger *et al.* 2023). Ernst *et al.* (2020) suggested Large Igneous Province (LIP) volcanism in the Siberian and Baltica Cratons dating between 380–360 Ma, which they implicated as triggers for the Kellwasser and Hangenberg Events. The Yakutsk-Vilui LIP shows peaks in magmatism at 374.1 (+/- 3.5 Myr) and at 363.4 (+/- 0.7 Myr), whereas the Kola Dnieper LIP is less accurately constrained (Ernst *et al.* 2020). The recorded peaks are only loosely correlated to the events in question: the first peak occurred at 2 Ma prior to the Kellwasser Event, whereas the second peak is 4.5 Myr prior to Hangenberg. Zhao *et al.* (2022) produced high resolution geochemical analyses of the extinction interval in China and North America, where they found that mercury isotopes in a correlative of the Upper Kellwasser horizon in China were consistent with a volcanic source. This signal is however, not recognised globally whereas, by contrast, other LIPs in the rock record associated with extinction horizons produced global mercury spikes (Zhao *et al.* 2022; Svensen *et al.* 2023). Such events would be expected to have produced isotopically light carbon, but the Kellwasser horizons record excursions in $\delta^{13}\text{C}$ and this too is inconsistent with a globally significant volcanic event (Zhao *et al.* 2022). Mercury enrichments preferentially accumulate in organic-rich marine sediments and may be derived from several

sources, such as enhanced continental weathering or wildfires (Liu *et al.* 2021; Pippenger *et al.* 2023). Percival *et al.* (2020) document phosphorous fluxes in several geographically widespread Upper Kellwasser Event horizons. According to their study the clearest explanation for phosphorous flux is increased soil development due to land vegetation, but a multi-causal trigger, including volcanism, orogeny and oceanic upwellings, was not ruled out (Percival *et al.* 2020). Similarly, Carmichael *et al.* (2019) favour a complex multicausal hypothesis, with a top down (i.e. continentally derived) mechanism for oceanic anoxia, particularly affecting restricted basins in tropical locations. Climate change related to astronomical forcing has been suggested based on cyclostratigraphy (De Vleeschouwer *et al.* 2017; Ma *et al.* 2022; Wichern *et al.* 2024). Astronomical forcing may have triggered the event due to a background of increasing soil development by land vegetation (Wichern *et al.* 2024).

2.1.2. The Hangenberg Event

The end-Devonian Hangenberg Event was a first-order extinction event comparable in impact to the ‘Big Five extinction events’, which caused a 50% loss of vertebrate species diversity (Sallan and Coates 2010) and severe extinctions amongst trilobites, ammonites and other marine invertebrate groups (Caplan and Bustin 1999; see Kaiser *et al.* 2016 for discussion on invertebrate extinctions), as well as floral turnover (Marshall *et al.* 2020). The crisis, as it is expressed in the terrestrial realm, indicated by overturns in miospore biodiversity and plant extinctions, occurred slightly later (100–300 kyr) than the marine extinctions (LN/VI zone boundary, Kaiser *et al.* 2016). The marine crisis interval is marked by widespread black shale deposits (the Hangenberg Black Shale) in marine basins (mostly below 40° latitude according to Caplan and Bustin 1999), generally reconstructed as a eustatic transgression immediately preceding the glacio-eustatic Hangenberg Sandstone regression (Caplan and Bustin 1999; Myrow *et al.* 2014; Kaiser *et al.* 2016; Denayer *et al.*

2021), though Bábek *et al.* (2016) preferred a regressive scenario for the black shale. There is disagreement about where in the sequence the main crisis occurs, some authors placing the crisis during the hypoxic Hangenberg Black Shale Event (Myrow *et al.* 2014; Kaiser *et al.* 2016) and others during the Hangenberg Sandstone regression (Becker *et al.* 2021).

Various triggering mechanisms are proposed; bolide impacts, atmospheric perturbations, volcanism, climatic cooling, climatic warming, oceanic overturns, eutrophication, sea-level changes, astronomical forcing, supernova explosions and evolution of land plants (Algeo and Scheckler 1998; Caplan and Bustin 1999; De Vleeschouwer *et al.* 2013; Kaiser *et al.* 2016; Fields *et al.* 2020; Rakocinski *et al.* 2021). Evidence for a bolide impact includes localised microtektites and concentrations of siderophile elements reported from south China (Ma and Bai 2002), with possibly contemporaneous impact craters in Quebec and Australia, however, lack of evidence for impact ejecta outside of China argues against this explanation for the Hangenberg Crisis (Caplan and Bustin 1999; Kaiser *et al.* 2016). Eutrophication and algal blooms during peak transgression and deposition of the Hangenberg Shale have been related to oceanic overturns (Kaiser *et al.* 2016) or, based on geochemical evidence, to increased terrestrial effluent entering marine basins (Carmichael *et al.* 2016; Percival *et al.* 2019). Plant spores from latest Famennian strata of the Old Red Sandstone Continent exhibited numerous malformations, and dark pigmentation (Marshall *et al.* 2020). Following experimental evidence, this was suggested to be the result of increased ultraviolet radiation due to the temporary loss of the ozone layer during a 9 kyr interval of rapid warming (Marshall *et al.* 2020). Malformed spores from other basins (i.e. Poland and Belgium) were considered to be an indicator of the Hangenberg Mass Extinction on land (Prestianni *et al.* 2016). An exploding supernova was proposed as a kill mechanism, but without present evidence to support the hypothesis (Fields *et al.* 2020). The volcanic hypothesis based on mercury

enrichments in the anoxic Hangenberg Black Shale continues to receive attention (Rakocinski *et al.* 2021; Smart *et al.* 2023).

2.1.3. Late Devonian biotic crisis

This brief review of ongoing debate on the causes of the Kellwasser and Hangenberg Events shows that no consensus is yet achieved for the cause of either event. Most extinction scenarios point to marine anoxia associated with rapid climate and sea-level changes. The intermittent marine anoxia and extinction scenario behind these and other ‘lesser’ events – i.e. the late Famennian Dasberg Crisis (Hartenfels and Becker 2009) and the Middle Famennian *annulata* Event (Korn 2004) – lends itself to a unifying explanation.

Bond and Wignall (2008) advanced the transgression–anoxia–extinction scenario based on a thorough investigation of the Kellwasser Event and comparison to other Late Devonian events. Geochemical evidence from the anoxic black shale event beds points to nutrient fluxes derived from continental rather than sea-floor derived sources (Carmichael *et al.* 2019; Percival *et al.* 2019). Plausible ‘top-down’ sources of this nutrient include volcanic activity and modification to the landscape by plants. The Devonian Plant Hypothesis proposes that Late Devonian forestation led to increased weathering of rock and soil development, leading to nutrient fluxes to the oceans and climate change associated with increased carbon burial (Algeo and Scheckler 1998; Algeo *et al.* 2001). With environmental instability due to this background of plant-based alterations to the carbon cycle, likely ultimate triggers for Late Devonian anoxic events appear to be volcanism (Smart *et al.* 2023) or climate change related to orbital cyclicity (De Vleeschouwer *et al.* 2013; Wichern *et al.* 2024).

2.1.4. Dating the Devonian-Carboniferous Boundary (DCB)

Problems with stratotype section for the Devonian-Carboniferous Boundary at La Serre, France have led to debates about where the boundary should be placed. Becker *et al.* (2016) reviewed the geological expression of the Hangenberg Event and the Devonian-

Carboniferous Boundary. Due to ongoing debates about the placement of the DCB they weighed the merits of various options, concluding that the Hangenberg Shale presented the greatest likelihood for global correlation, however, cautioning that this horizon may, in some sections, have been removed by subsequent regression. The DCB is presently recognised at within the upper Hangenberg Crisis interval overlying the Hangenberg Sandstone, thought to represent a postglacial transgression, by the first appearance of the conodont *Siphonodella sulcata* (Becker *et al.* 2020). An average of various estimates for this biostratigraphic horizon gives a date of 359.3 ± 0.3 Ma (Becker *et al.* 2020), which is slightly older than the previously accepted date of 358.9 Ma (Becker *et al.* 2012; Cohen *et al.* 2013). More recently, researchers from the International Working Group on the DCB proposed a globally correlatable eustatic sequence boundary, the top of the Hangenberg Sandstone, to mark the DCB (Aretz and Corradini 2021; Denayer *et al.* 2021).

2.2. Geological setting

The Cape Supergroup contains an up to 8 km thick siliciclastic sedimentary succession deposited from the Ordovician Period until the Early Carboniferous Period in a passive margin extensional basin (for detailed review see Broquet 1992; Shone and Booth 2005; Thamm and Johnson 2006). It unconformably overlies a Precambrian to Cambrian basement of deformed metasediments and granitic intrusions and is para- to unconformably overlain by the Karoo Supergroup (Thamm and Johnson 2006). Strata of the Cape Supergroup outcrop in the Cape Fold Belt, occupying a zone within 200–250 km of the coast in the Eastern and Western Cape provinces, extending a further 100–200 km inland in the subsurface and approximately 100 km offshore (Tankard *et al.* 2012). The strata were compressed and metamorphosed during a regional Late Palaeozoic orogenic event that formed the Gondwanide mountains, evidenced by contemporary fold and thrust belts in the Cape, Antarctica and South America (Miller *et al.* 2016). Subsequent extension during the

Mesozoic breakup of Gondwana initiated northwest-southeast and east-west trending normal faults that further fragmented the Cape Supergroup (Booth 2011). The Cape Fold Belt comprises two arms, an east-west trending southern arm, characterised by intense shortening (up to 70%) and a north-south trending western arm with a lesser degree (~20%) of shortening (Shone and Booth 2005; Paton *et al.* 2006; Booth 2011). These meet and interfere as a syntaxial oroclinal bend in the Ceres area, marked by northeast trending folds (de Beer 1995). The structural strike of the Cape Fold Belt controls the present-day orientation of the South African coastline. Structural deformation in the Cape Fold Belt reaches its highest intensity in the Eastern Cape, where the metamorphic grade is lower Greenschist facies (around 300°C) (Paton *et al.* 2006). This eastward intensity of deformation has caused the thrust stacking and thickening of all stratigraphic units (Booth *et al.* 2004; Booth 2011). Booth (2011) compared two tectonic models that he considered most consistent with the characteristics of this fold belt. The Andean-type model infers a subduction zone around 1 500 km to the south of the fold belt (Booth 2011). Interference folding between the western and southern limbs in a region known as the syntaxis is explained by the presence of rotating microplates (de Wit and Ransome 1992). The Andean-type model does not adequately explain the ‘thin-skinned’ structural characteristics observed in parts of the fold belt (Stankiewicz *et al.* 2007; Booth 2011), and many typical features of a subduction zone such as ophiolites, widespread volcanism and high-grade metamorphism are lacking (Newton *et al.* 2006). The transpression model proposed by Johnston (2001) invokes right-lateral strike slip motion along a plate boundary to the south. This model is consistent with the asymmetry of the southern and western arms of the Cape Fold Belt as well as broader evidence from contemporary orogenic belts in Argentina and Antarctica, although several structural anomalies have not been reconciled with the model (Booth 2011). Deformation of the Cape Fold Belt led to subsidence in the continental interior to the north, forming a continental

foreland basin that accommodated the Permian to Jurassic Karoo Supergroup sediments and lavas (McCarthy and Rubidge 2005).

The Witteberg Group, which is the upper unit of the tripartite Cape Supergroup, is exposed along most of the extent of the Cape Fold Belt, from the Great Fish River in the Eastern Cape to Elandsvlei (about 60 km WSW of Clanwilliam) in the Western Cape where it is erosionally truncated by the Dwyka Group (Loock 1976). It represents a continuous record of siliciclastic shallow-marine to continental sedimentation from the Givetian (Middle Devonian) until the Visean (Early Carboniferous) (Broquet 1992; Shone and Booth 2005; Thamm and Johnson 2006). A correlative unit to the Witteberg Group, which was not incorporated in the Cape Fold Belt, is the Msikaba Formation that outcrops along the border of KwaZulu-Natal and the Eastern Cape (Kingsley and Marshall 2009). The Msikaba Formation is thought to be Late Devonian in age based on plant remains from one horizon in the formation (Anderson and Anderson 1985).

Johnson (1991) proposed that Witteberg Group sandstone (as all Cape Supergroup sandstone) was derived from a cratonic interior source based on the supermature quartzitic lithologies that predominate, and suggested a low-relief continental source to the north. Proterozoic ages for the source rocks derived from detrital zircon studies suggested the Namaqua–Natal Metamorphic Province and Pan-African Orogeny as likely source areas (Vorster 2013). The sediments accumulated within the Agulhas Sea, a passive-margin marine basin to the south of Gondwana, which was envisaged by Marshall (1994) and Rust (1973) to have been semi-enclosed from the open ocean by the Patagonian and Antarctic Peninsulas.

Witteberg Group stratigraphy was formalised in the South African Commission for Stratigraphy in 1980 (SACS 1980) and later revised by Johnson *et al.* (2006). It contains three subgroups (each consisting of three formations) and one formation. Lithologically the Witteberg Group is between 1 200 m (Western Cape) and 1 700 m (Eastern Cape) thick

(Johnson *et al.* 2006) and is suggested to contain sandstone and mudstone in roughly equal proportions (Johnson 1976). It conformably overlies shallow-marine deposits of the Bokkeveld Group, and is unconformably overlain by glacial tillite of the Dwyka Group (Dunlevey and Hiller 1979; Johnson *et al.* 2006). In the Eastern Cape, however, the Witteberg–Dwyka contact is possibly conformable (Loock 1967).

The stratigraphic hierarchy that is presently used for the Cape Supergroup and its subdivisions was first applied by Johnson (1966), based on recommendations by the International Subcommission on Stratigraphic Classification. This system was revised by Johnson (1976), and the stratigraphic framework for the Witteberg Group there applied remains little changed. The historical development of Witteberg Group stratigraphy prior to this was reviewed by Loock (1967), who credits its first recognition as a distinct stratigraphic unit to A.G Bain's 1856 geological map. Loock (1967) documents the progress in the formalising of this stratigraphy beginning around the start of the 20th Century when the Geological Commission of the Cape of Good Hope employed geologists, such as G.S. Corstorphine, E.H.L. Schwarz, and A.L. Du Toit. Separating the Witteberg Group from the Bokkeveld Group has long challenged geologists due to the continuity of sedimentary style (Swart 1950; Theron 1960). According to Johnson (1976), the Witteberg Group is distinguished from the Bokkeveld Group by several characteristics, including an abundance of trace fossils (*Spirophyton* in particular), the ubiquity of upward-coarsening cycles, the micaceous nature of sediments, the abundance of plant stems and the reddish colouration of rhythmites (heterolithic units) (Johnson 1976). Haughton (1969) placed the Bokkeveld–Witteberg boundary at the first prominent white-weathering sandstone unit, which was supported by Johnson (1976) and Hiller and Dunlevey (1978) who recognised this horizon as the base of the Nougaspoort Member (Wagen Drift Formation). Theron (1972) suggested that the Witteberg Group was better distinguished by a conspicuous increase in the heavy mineral

Brookite, and proposed a stratigraphically higher sandstone unit, however, the Nougaspoot Member was formally selected as the base of the Witteberg Group (Johnson *et al.* 2006).

There was much disagreement, prior to Johnson (1976), about the position of the Witteberg Group's upper contact. Historically, the strata overlying what is today the Floriskraal Formation (Lake Mentz Sub-Group) were named the 'Lower Dwyka Shales', however, du Toit (1926) and Rossouw *et al.* (1953) included a portion of these units (Waaipoort Formation) in the 'Upper Witteberg Shales'. Rossouw placed the 'basal tillite' (now the Miller Formation) within the overlying Dwyka Series, placing the Cape–Karoo Systems' contact at what is today the base of the Kommadagga Subgroup, an approach that was supported by Loock (1967). Loock regarded the 'Kommadagga Formation' to be conformable with the overlying 'Dwyka Tillite Formation' in the Eastern Cape, whereas this unit is erosionally truncated to the west by a subglacial unconformity at the base of the Dwyka Tillite (Dunlevey and Hiller 1979). The Kommadagga was raised to Subgroup status by Johnson (1976) and is included in the Witteberg Group (Johnson *et al.* 2006), placing the Cape Supergroup–Karoo Supergroup boundary at the base of the Dwyka Group.

2.2.1. *The lower Witteberg Group*

The lower Witteberg Group includes the Weltevrede Formation in the Eastern Cape (east of 23°E) and the Weltevrede Subgroup in the Western Cape, which is divided into three formations (Wagen Drift, Blinkberg and Swartruggens). The Blinkberg Formation in the Western Cape is correlated with a Blinkberg Member in the Eastern Cape (Hiller 1990). Based on the Early Devonian marine invertebrate faunas of the lower Bokkeveld Group, Theron (1972) suggested that the Witteberg Group was Late Devonian to Early Carboniferous in age. In agreement with this, the Weltevrede Formation/Subgroup was indicated to predate the Famennian Stage based on occurrences of the articulate brachiopod *Tropidoleptus* near the base and top of the unit (Boucot *et al.* 1983). By comparison with

Bolivian occurrences of the genus, a Givetian to Frasnian age was considered most likely (Boucot *et al.* 1983). Cooper (1986) considered the black shale horizons in the upper member of the Wagen Drift Formation (Bynest Krans Member) to indicate a maximum flooding surface, and thus correlated this event to T–R cycle IIa of Johnson *et al.* (1985), indicating a mid-Givetian age for the Wagen Drift Formation. Shallow to marginal-marine depositional environments ascribed to the lower Witteberg Group include deltaic deposits (Theron and Looock 1988; Hiller 1990) and storm-dominated shallow-marine deposits (Cole and Labuschagne 1983; Cotter 2000). Much like the Bokkeveld Group, a proximal-distal gradient from north to south is evidenced by flow vectors, isopach maps and the southward attenuation of sandy units (Theron and Looock 1988; Cotter 2000). The lower Witteberg Group is considered to represent shallowing of the basin with respect to the Bokkeveld Group (Theron and Looock 1988). Hiller (1990) described various deltaic sub-environments for the uppermost strata of the Weltevrede Formation near Makhanda (Eastern Cape), recognising interdistributary bay fill deposits, prograding distributary mouth bar deposits and tidal flats. Cole and Labuschagne (1983) made a detailed facies analysis of a heavy mineral placer deposit in the upper part of the Weltevrede Formation near Willowmore, 100 m below the contact with the Witpoort Formation, which they interpreted as a marine tempestite deposited below fair-weather wave base. Likewise, Cotter (2000) favoured a storm-dominated shallow-marine model. The Weltevrede Subgroup, according to Cotter (2000) comprises approximately 40 upward-shallowing parasequences averaging 12 m thick, representing inner-shelf to shoreface deposits.

By comparison to the Bokkeveld Group there is a general paucity of marine invertebrate remains in the Lower Witteberg Group. Rare marine invertebrate assemblages in the Wagen Drift Formation include a low diversity of articulate and inarticulate brachiopods and bivalves (Boucot *et al.* 1983). These groups, as well as a trilobite and a gastropod, were

identified from the uppermost strata of the Weltevrede Formation (Hiller 1990). A poorly preserved, putative conchostracan (Rennie 1934), which was later identified as a bivalve, possibly *Naiadites devonicus* (Scholze and Gess, 2021), as well as fragments of a eurypterid identified as *Hasimima* by Woodward (1909), were found in the vicinity of Port Alfred, in strata that Rennie and Mountain (1942) included in the upper Bokkeveld Series, which were subsequently mapped in the lower Witteberg Group (Booth *et al.* 1999), but may have come from the Witpoort Formation (Scholze and Gess 2021). Plant remains and a diversity of trace fossils occur throughout the lower Witteberg Group (Hiller 1990). The trace fossils have not been described in detail, with the exception of *Spirophyton* described by Plumstead (1967), who considered the trace to be an aquatic plant. Almond (1998) reported from the Weltevrede Subgroup, a low to moderate diversity of trace fossils including *Zoophycos*, *Teichichnus*, *Scolicia* Group meniscate burrows, lingulid brachiopod burrows, *Palaeophycus*, *Diplo craterion* and several unnamed forms. Plant remains were described by Plumstead (1967), but her results are of limited stratigraphic utility because she included the lower Witteberg Group strata in the Bokkeveld Group. Anderson and Anderson (1985) reviewed the Cape Supergroup flora with an up-to-date stratigraphy. They identified enigmatic plant taxa *Dutoitia alfreda*, *Palaeostigma sewardi* and the lycopods *Haplostigma irregularis*, *Haplostigma kowiensis* and *Archaeosigillaria caespitosa* from the upper Weltevrede Formation, whereas the lower Weltevrede Formation contains only *Haplostigma* and rare *Archaeosigillaria*. Gess (2016) reviewed the vertebrate fossil record of the Witteberg Group, noting, from the Weltevrede Formation, only fragmentary fish remains from a lag deposit in the Wagen Drift Formation, which he identified as shark spines of the *Plesioselachus*-type and arthrodire placoderm remains reminiscent of *Barrydaleaspis*.

2.2.2. *The Witpoort Formation*

The Witpoort Formation forms the middle unit of the Witteberg Group (Johnson *et al.* 2006). It conformably overlies the Weltevrede Formation with a sharp or gradational contact, that, according to Johnson (1976), is sharper eastwards in the basin. Johnson (1976) defines the lower boundary of the Witpoort Formation as the horizon above which sandstone is conspicuously more abundant than mudstone or streaky rhythmite. The upper contact of the Witpoort Formation is described as conformable and sharp or intertongued, with a transition zone of a few metres thick (Johnson 1976). According to Loock (1967) this Witpoort Formation to Lake Mentz Subgroup transition is usually not well exposed but, where visible, it is a 25 feet (7.6 m) thick alternation of sandstones, siltstones and micaceous shales. The contact is defined as the base of the shale-dominated Kweekvlei Formation (lowermost formation of the Lake Mentz Subgroup) (Johnson 1976). Prior to Johnson (1976), the Witpoort Formation had been referred to as the Main Stage of the Witteberg Series by Theron (1960), and the Wittepoort Formation by Loock (1967), named after a locality approximately 30 km NW of Grahamstown. Since the formation is only partially exposed at Wittepoort, the name was revised to Witpoort, after a section north of the Gamkapoort Dam near Prince Albert, Western Cape, where the formation is fully exposed, and which defines the stratotype section (Johnson 1976; Theron *et al.* 1991). Near its westernmost extent the Witpoort Formation is as little as 140 m (Theron and Loock 1988). In the central basin around the Eastern Cape/Western Cape border, between 23°E and 25°E, thickness of the Witpoort Formation is approximately 300 m (Johnson 1976). A stratigraphic log from within this zone gives a thickness of approximately 350 m at Waaipoort, near Steytlerville (Johnson *et al.* 2006). Based on thickness figures of 850 m and 885 m in an unpublished report by D.K. Toerien, 730 m by Mountain (1946) and 1200 m by Meyer (1965), Johnson (1976) estimated the Witpoort Formation to be 800 m thick in the Grahamstown area. Johnson considered

Meyer's (1965) suggestion that 'drag folding' was responsible for abnormal thicknesses of the Witpoort Formation in this area to be implausible. Hiller and Taylor (1992), however, estimated the Witpoort Formation to be approximately 350 m thick in the same area. Following Johnson's (1976) PhD results (with no new stratigraphic data), Johnson *et al.* (2006, Table 1) gave a thickness of 850 m eastward of 22°E.

The Witpoort Formation is a sandstone-dominated succession that is separated into three members (Rooirand Member, Perdepoort Member and Skitterykloof Member) westward of 25° E (SACS, 1994; Thamm and Johnson 2006). Further east, however, the formation is not formally subdivided (Hiller and Taylor 1992). The Rooirand Member is between 250–300 m thick (SACS 1994) and forms the bulk of the Witpoort Formation's thickness. Its name is derived from the Rooirand Hills, west of the Gamkapoort Dam near Prince Albert (Western Cape), where J.C. Loock placed the stratotype. The Rooirand Member typically comprises brown to red-weathering, fine to medium-grained quartzitic sandstone forming lenticular or tabular beds averaging 30 cm in thickness, with poor preservation of internal sedimentary structures (Johnson 1976). A minor proportion of mudrock and 'streaky rhythmitite' (6%) is recorded in the Rooirand Member at Waaipoort (Eastern Cape) (Johnson 1976). It is conformably overlain by the Perdepoort Member, a white-weathering quartzitic sandstone unit that is 30–60 m thick. It was originally known as the White Streak ('Witstreep') (Loock 1967) due to a distinctive whitish colour in outcrop visible in aerial photographs. It is characterised as being composed exclusively of mature, clean quartzitic sandstone exhibiting very regular tabular strata with horizontal and low angle cross-stratification (Johnson 1976). The name of this member was introduced by Johnson (1976), derived from Perdepoort north of Willowmore (Eastern Cape), where a quarry exposes the entire member.

The Skitterykloof Member conformably overlies the Perdepoort Member, occurring sporadically at the top of the Witpoort Formation (Johnson 1976; Thamm and Johnson 2006).

It comprises massive sandstone and a variable proportion of mudrock (10–40% according to Johnson 1976). It is characterised by the presence of coarse-grained, pebbly, immature sandstone and conglomerate, distinguishing it from the underlying Perdepoort Member (SACS 1994; Thamm and Johnson 2006). The Skitterykloof Member is conformably overlain by the Kweekvlei Formation. The Skitterykloof Member was named by J.C. Looock after a locality northeast of Ceres (Western Cape) where the whole member is exposed. An approximate thickness of 15 m is shown in a stratigraphic log (Fig. 2.1A) derived from Waaipoort (Johnson *et al.* 2006).

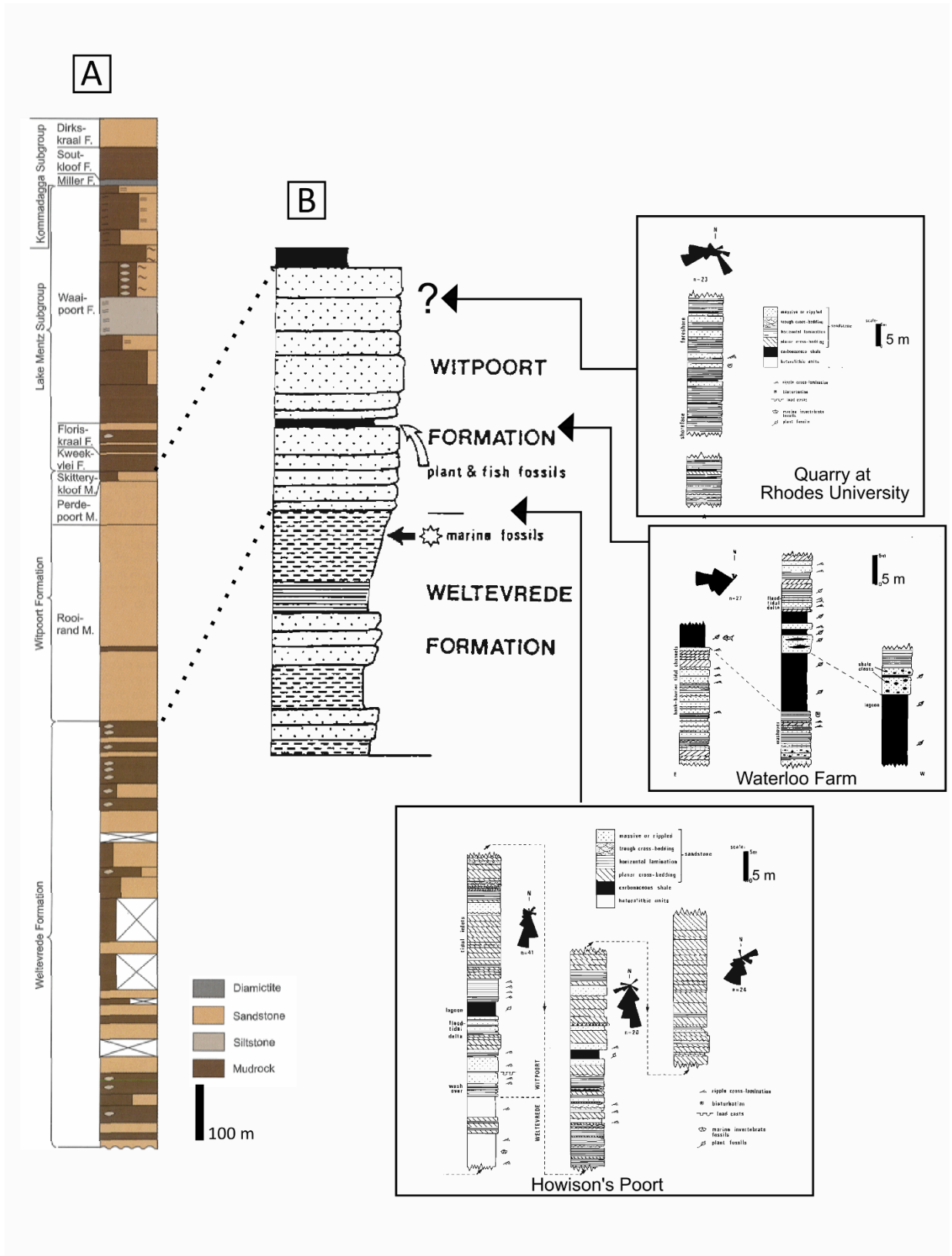


Figure 2.1: Stratigraphic columns for the study area: A) The Witteberg Group, logged at Waaipoort (Johnson *et al.* 2006). B) The eastern sector (Hiller and Taylor) with enlarged metre-scale columns for three localities in the vicinity of Grahamstown, showing their approximate stratigraphic positions.

The transition from the Weltevrede Subgroup/Formation to the Witpoort Formation has been suggested to have resulted from a change in shoreline configuration from lobate to linear (Johnson 1976; Theron and Looock 1988; Hiller 1990; Hiller and Taylor 1992) and represents a shallowing event (Cotter 2000) that is related to a reduction of subsidence in the basin (Hiller and Taylor 1992) or falling sea-levels (Cooper 1986). A linear east-west trending high-energy shoreline is widely accepted for the genesis of the supermature sandstone-dominated Witpoort Formation (Theron and Looock 1988; Hiller and Taylor 1992; Thamm and Johnson 2006). Westward-directed palaeocurrent signals in the formation are thought to have been generated by shore-parallel currents (Looock 1967; Theron and Looock 1988; Johnson 1991; Thamm and Johnson 2006). Based on these inferred shore-parallel palaeocurrent signals a shoreface setting for the Rooirand Member in the central part of the basin was proposed by Thamm and Johnson (2006), whereas further east, southwest directed palaeocurrents potentially indicate a fluvio-deltaic influence in this member (Thamm and Johnson 2006). The overlying Perdepoort Member is interpreted as a beach deposit based on the predominance of low angle cross-stratification and the great lateral extent of strata (Theron and Looock 1988; Thamm and Johnson 2006). In the eastern part of the basin near Makhandia, where the three members of the Witpoort Formation are not recognised, Hiller and Taylor (1992) identified a barrier island setting, represented by shoreface, beach, tidal inlet, tidal delta, washover fan, back-barrier tidal channel, and lagoonal deposits (Fig. 2.1B). Carbonaceous shale developed in stagnant brackish water lagoons which accumulated and buried abundant terrigenous plant and marine algal matter (Hiller and Taylor 1992), in counter to the hypothesis that the muds were deposited offshore (Johnson 1976). At the nearby Coombs Hill fossil locality, an approximately 80 m thick stratigraphic section of the middle (?) Witpoort Formation was interpreted as containing shoreface, beach and lagoon deposits (Harris *et al.* 2021).

Clast-poor sandy diamictite from the upper stratigraphic horizons of the Witpoort Formation, including above and below the Perdepoort Member, were interpreted as glacial deposits thought to correspond to late or latest Famennian glaciation events recorded in Bolivian strata (Almond *et al.* 2002).

Plumstead (1967) suggested that *Leptophloeum* fossils from this ‘Main Witteberg Quartzite’ indicated a Middle to Late Devonian age. In contradiction of the above, Rayner (1988) proposed an Early Devonian age for the Witpoort Formation based on ‘psilophyte’ plant remains described by Plumstead (1967). The discovery of *Archaeopteris* confirmed that the Witpoort Formation was Late Devonian (Hiller and Taylor 1992; Anderson *et al.* 1995). Cooper’s (1986) event stratigraphy model correlated the Witpoort Formation with Johnson *et al.*’s (1985) T–R cycles IIe and IIf, representing regressive events during the Famennian stage. Cooper (1986) proposed that a flooding event at the Witpoort–Kweekvlei Formation (lower Lake Mentz Subgroup) contact represented an early Tournaisian transgression recorded in the Rhenish Slate of Germany, thus placing the Devonian–Carboniferous boundary at approximately this horizon. The strongest evidence supporting a Famennian age for the Witpoort Formation comes from macrofossils derived from one stratigraphic horizon at one locality, whereas microfossils are next to unknown in the unit. Miospores indicating a Famennian age were reported from a horizon underlying the Perdepoort Member, but without identification or description of the spores (Almond *et al.* 2002). Fossil fauna and flora from Waterloo Farm include various taxa with a Late Devonian signal. The locality is considered to be situated in the uppermost Witpoort Formation (Gess 2011; Scholze and Gess 2021). Plant genera such as *Archaeopteris* and *Leptophloeum* are typical for the Late Devonian (Anderson *et al.* 1995; Osborne *et al.* 2004; Wang *et al.* 2005; Cornet *et al.* 2012). Undescribed putative spermatophyte remains from the locality (Gess and Whitfield 2020) would strongly indicate a Famennian age (Prestianni *et al.* 2015). Vertebrates including

tristichopterids (*Hyneria udlezinye*) and stem tetrapods (*Tutusius umlambo* and *Umzantsia amazana*) from the Waterloo Farm fossil locality likewise indicate a Late Devonian age (Gess and Ahlberg 2018, 2023). Apart from its occurrence at Waterloo Farm, *Hyneria* is only known from one other locality, Red Hill in the Catskill Formation of the eastern USA, which is late Famennian in age (Broussard *et al.* 2018). Placoderms are diverse and abundant at Waterloo Farm, but do not occur in stratigraphically higher strata (Gess 2016). Placoderm extinction at the Devonian–Carboniferous boundary (Sallan and Coates 2010) is consistent with the current recognition of this boundary at the top of the Witpoort Formation (Gess 2016). The anoxic shales of the overlying Kweekvlei Formation are, however, very poor in fossils and therefore the absence of placoderms or other Devonian taxa in this formation does not prove an Early Carboniferous age. According to Gess (2016), the paucity of vertebrate remains in this stratigraphic interval is explained with reference to Romer’s Gap, a 15–20 Myr interval spanning the Tournaisian Stage (earliest Carboniferous), during which vertebrate and arthropod fossil remains are uncommon (Ward *et al.* 2006; Sallan and Coates 2010). Some, however, consider Romer’s Gap to be an artefact of sampling bias (Byrne *et al.* 2022). Palaeontological evidence therefore appears to support Cooper’s (1986) event stratigraphy correlation of the Witpoort Formation with the Famennian stage (Theron 1994; Gess 2016).

Palaeobotany of the Witpoort Formation was reviewed by Plumstead (1967), Anderson and Anderson (1985) and Gess (2011). Anderson and Anderson (1985) performed a taxonomic and stratigraphic revision of Plumstead’s work, including altering the names of taxa, performing synonymies, and introducing new taxa. These revisions are explained in detail by Gess (2011). Anderson and Anderson (1985) recognised nine plant taxa in the Witpoort Formation including four lycopods (*Archeosigillaria caespitosa*, *Leptophloeum australe*, *Palaeostigma robusta* and *Palaeostigma sewardi*), a ‘psilophyte’ (*Dutoitia*

maraisia) and three plants of uncertain affinity (*Howisonia rara*, *Platyphyllum albanense* and *Praeramunculus alternatiramus*). Apart from *A. caespitosa* and *L. australe*, all of these plant species were derived from a thin black shale horizon in the lower Witpoort Formation at Howison's Poort.

Subsequent to Anderson and Anderson's review, several plant and algae taxa have been diagnosed from the Witpoort Formation, principally from one stratigraphic horizon at Waterloo Farm. Plants from Waterloo Farm include phaeophyte or rhodophyte alga, *Hungerfordia fionae* and *Yeaia africana* (Hiller and Gess 1996), charophyte algae comprising two genera and four species, *Octochara crassa*, *O. gracilis*, *Hexachara setaceae* and *H. riniensis* (Gess and Hiller 1995b), three lycopods, *Leptophloeum rhombicum* (Prestianni and Gess 2014), *Kowieria alveoformis* (Gess and Prestianni 2018) and *Colpodexylon pullumpedes* (Harris *et al.* 2021b), a fern-like plant of uncertain higher affinity, *Flabellopteris lococannensis* (Gess and Prestianni 2022), a sphenophyte, *Rhinistachya hilleri* (Prestianni and Gess 2018) and a progymnosperm, *Archaeopteris notosaria* (Anderson *et al.* 1995; Gess and Berry 2024). One plant species was diagnosed from Coombs Hill, a lycopod *Colopdexylon mergae* (Harris *et al.* 2021b). Additional undescribed plant remains from Waterloo Farm include zosterophylls and seed plants (Gess and Whitfield 2020).

Described invertebrate diversity in the Witpoort Formation is limited to a monospecific assemblage of bivalves, *Naiadites devonicus* (Scholze and Gess 2021) from Waterloo Farm, representing the oldest record of a nonmarine genus that became widespread in the Carboniferous Period (Scholze and Gess 2017) and remains of a terrestrial scorpion *Gondwanascorpio emzantsiensis* (Gess 2013). Undescribed invertebrates include remains of a cyrtoctenid eurypterid (Gess and Hiller 1995a; id. by R.W. Gess 2021), possible ostracods (Scholze and Gess 2021) as well as various bivalve species from Coombs Hill (Harris *et al.*

2021a). Lingulid brachiopods (*Dignomia* cf. *lepta*) are reported and described in this thesis (Chapter 7) from a lower Witpoort Formation locality (Harris and Gess 2022).

Vertebrate remains in the Witpoort Formation are almost exclusively reported from Waterloo Farm. They include a lamprey, *Priscomyzon riniensis* (Gess *et al.* 2006), with an ontogenetic series that includes hatchlings to adults (Miyashita *et al.* 2021), placoderms, including an antiarch, *Bothriolepis africana* (Long *et al.* 1997), and three arthrodire species, *Groenlandaspis riniensis* (Long *et al.* 1997), *Africanaspis doryssa* (Long *et al.* 1997) and *Africanaspis edmountaini* (Gess and Trinajstić 2017), the acanthodians *Diplacanthus acus* (Gess 2001) and *Gyracanthides riniensis* (Gess and Burrow 2024), three chondrichthyans *Antarctilamna ultima* and *Plesioselachus macrocanthus* (Gess and Coates 2015b) and an undescribed symmoriid (Gess 2024), an actinopterygian (Gess and Whitfield 2020; Choo and Gess, in prep) and various lobed-fin fishes; a coelacanth *Serenichthys kowiensis* (Gess and Coates 2015a), a lungfish *Isityumzi mlomomde* (Gess and Clement 2019), a tristichopterid *Hyneria udlezinye* (Gess and Ahlberg 2023), an undescribed rhizodont (Gess and Whitfield 2020), an underscribed onychodont (Gess 2011) and two tetrapod species, *Umzantsia amazana* and *Tutusius umlambo* (Gess and Ahlberg 2018). Reviewing the palaeontology of the Waterloo Farm locality, Gess and Whitfield (2020) proposed an estuarine environment by analogy to modern and ancient estuarine ecology, including the occurrence of marine and nonmarine fish and algal species, evidence for piscean utilisation of the environment as a nursery, and the predominance of a single shelly invertebrate species. Only one other vertebrate fossil is reported from the Witpoort Formation, an arthrodire placoderm skull-roof (Harris and Gess 2022), which is from a lower Witpoort Formation locality described in this thesis (Chapter 7).

Trace fossils from the Witpoort Formation are to be found in very few reports. *Spirophyton/Zoophycos* is indicated to occur in the Witpoort Formation in various reports

(Plumstead 1967; Looock 1967; Gess and Hiller 1995a; Almond 1998), however, the distribution of the trace in the formation is difficult to discern. Plumstead (1967) records the trace as ranging from the lower Bokkeveld Series to the top of the Main Sandstone of the Witteberg Series without providing evidence for this. None of the specimens that she described are unequivocally from the Witpoort Formation. Looock (1967) cites Theron (1960) as recording the trace fossil *Spirophyton* ranging into the lower beds of the Witpoort Formation, but the highest stratigraphic record of the trace was within a 400 ft thick heterolithic unit underlying the Main Witteberg Quartzite (Theron 1960). Gess (2011) observed that *Spirophyton* frequently occurs in black carbonaceous shales near the bottom of the Witpoort Formation, and illustrated examples of the trace from the Waterloo Farm black shale. The trace fossils *Cruziana* and *Rusophycus* were identified from heterolithic sandstone underlying the shale horizon (Gess and Hiller 1995a). Hiller and Taylor (1992) recorded bioturbation in various stratigraphic logs from the Witpoort Formation, but did not describe the nature of the trace fossils.

Vorster (2013, her Fig. 3.8) illustrated a trace fossil surface with indeterminate vertical burrows in the Witpoort Formation near Laingsburg. Brunsdon (2006) illustrated a vertical and a horizontal burrow from the Perdepoort Member.

2.2.3. *The upper Witteberg Group*

The upper Witteberg Group includes the Lake Mentz Subgroup and the Kommadagga Subgroup (Johnson *et al.* 2006). The Lake Mentz Subgroup, which conformably overlies the Witpoort Formation, records the initial deepening of the basin (Kweekvlei Formation) followed by shallowing and outbuilding of a sandy shoreline (Floriskraal Formation), and a subsequent deepening indicated by the Waaipoort Formation (Broquet 1992). Marine and nonmarine environments are indicated for the Lake Mentz Subgroup (Broquet 1992).

The Kweekvlei Formation (25–200 m thick) is a mudrock-dominated succession with subordinate fine-grained sandstone units exhibiting hummocky cross-stratification, interpreted as storm influenced marine shelf sediments (Browning and Reid 2017). Fossils in the Kweekvlei Formation include indeterminate plant axes (e.g. *Praeramunculis* sp.) and trace fossils (*Teichichnus*-dominated, with rare *Cruziana* and *Palaeophycus*) (Browning and Reid 2017). Gess (2016) reported a shark spine from a conglomeratic horizon in the lowermost Kweekvlei Formation, but whether this horizon belongs to the uppermost Witpoort Formation needs to be reinvestigated.

The Kweekvlei coarsens upwards into the arenaceous deposits of the Floriskraal Formation, which is 70–130 m thick (Browning and Penn–Clarke 2016). The Floriskraal Formation is considered to represent a storm-dominated shallow-marine setting (Browning and Penn-Clarke, 2016). It contains a low diversity of horizontal burrows (Almond 1998) and poorly preserved plant remains, including lycopods (Browning and Penn-Clarke 2016). Possible fish fragments were reported from phosphatic nodules in the lowermost strata of the Floriskraal Formation (Evans 2005).

The overlying, up to 340 m thick, mud-dominated Waaipoort Formation has been interpreted as an extensive lacustrine (Evans 2005) or lagoonal deposit (Broquet 1992). The Waaipoort Formation contains the large sweep-feeding eurypterid *Cyrtoctenus wittebergensis* (Waterston *et al.* 1985), plant remains, bivalves and fishes (Evans 2005). The latter include acanthodians (cf. *Gyracanthides*), chondrichthyans (*Plesioselachus*-like), actinopterygians (seven genera and 11 species) and, possibly, sarcopterygians (Evans 1999; Gess 2016). Ichnofossils from this formation include *Teichichnus*, small *Cruziana* and *Rusophycus* of the *Isopodichnus* type, *Lockeia*, horizontal burrows and arthropod trackways (Almond 1998). Trace fossils in the Waaipoort Formation (and Lake Mentz Subgroup as a whole) indicate a nonmarine environment for these strata (Almond 1998).

Plumstead (1967) suggested a Late Devonian age for the Upper Witteberg Shales (now Waaipoort Formation, Lake Mentz Subgroup) based on fossil plant (lycopod) axes that she identified as *Protolepidodendron eximum* (now considered an invalid genus, Bonamo *et al.* 1988). This was supposedly supported by palynomorphs indicating a Middle to early Late Devonian age (Stapleton 1977). In contrast, Gardiner (1969) described a diverse fauna of palaeoniscoid actinopterygians which exhibited a great degree of higher taxonomic overlap with Lower Carboniferous (Visean) strata of the Cementstone Group of Scotland, in which six out of eight families from the Waaipoort Formation also occur, with the remaining two being endemic (Gardiner 1969). Streel and Theron (1999) refuted Stapleton's palynological result, producing miospores from the Waaipoort Formation that indicated a middle Tournaisian to early Visean age, in support of Gardiner's conclusion.

The Kommadagga Subgroup is considered to contain evidence for the earliest phase of glaciation of the Permo-Carboniferous ice age (Theron 1994). The basal Miller diamictite, interbedded with the Swartwaterspoort sandstone, provides evidence for overriding glaciers, and gives way to laminites of the Soutkloof Fm, and then to the Dirkskraal sandstone, which have been interpreted as a nearshore to deltaic deposit (Swart, 1982; Theron 1994). The Kommadagga Subgroup is considered to be largely unfossiliferous, except for some horizontal burrows of an unknown type from the Dirkskraal Formation (Almond 1998). The Kommadagga Subgroup is overlain by the Dwyka Group, representing the Cape Supergroup-Karoo Supergroup contact. The base of the Dwyka Group is thought to be Late Carboniferous to Early Permian in age (Theron 1994). It is generally agreed that a significant hiatus in deposition occurred prior to deposition of the Dwyka Group (Dunlevey and Hiller 1979; Swart 1982; Theron 1994).

CHAPTER 3 – MATERIAL AND METHODS

3.1. Material:

3.1.1. Study area

At the beginning of this study, its scope was determined by visiting 13 Witpoort Formation localities along the extent of the Cape Fold Belt. From these, seven in the Eastern Cape were selected for stratigraphic and sedimentological analysis. The criteria for selecting lithological sections were, firstly, anchorage to stratigraphic contacts, in order to circumvent the problem of floating sections that has affected stratigraphic work in the study area (Hiller and Taylor 1992; Harris *et al.* 2021a). Outcrops that presented continuous sections to recognisable formational or member boundaries were favoured. Intense structural deformation was considered a negative criterion. The second criterion was the presence of body and trace fossils, which would provide invaluable palaeoenvironmental information. For this reason, six out of seven studied sections are located within a 30 km radius of Makhanda (Eastern Cape), from which most of the Witpoort Formation palaeontological record is derived. This part of the study area will be referred to as the ‘eastern sector’ throughout this work. Approximately 200 km west of Makhanda, at Waaipoort (near Steytleville), a well exposed cross-section of the Witpoort Formation was selected as a reference. The study area includes selected sections of the Witpoort Formation in the Eastern Cape (Fig. 3.1). The sections were investigated during three annual field seasons from 2021 to 2023. The Witpoort Formation type section at Gamkapoort Dam (near Prince Albert, Western Cape, Theron *et al.* 1991) was eliminated from the study because the outcrop was inundated by an unusual rise in the dam level in the spring of 2022. Each studied section is presented with a map of approximately 1:10 000 scale with stereonet projections of the bedding orientations and structures recorded therein. Metre-scale stratigraphic logs for the sections are provided in Appendix 1C–L.

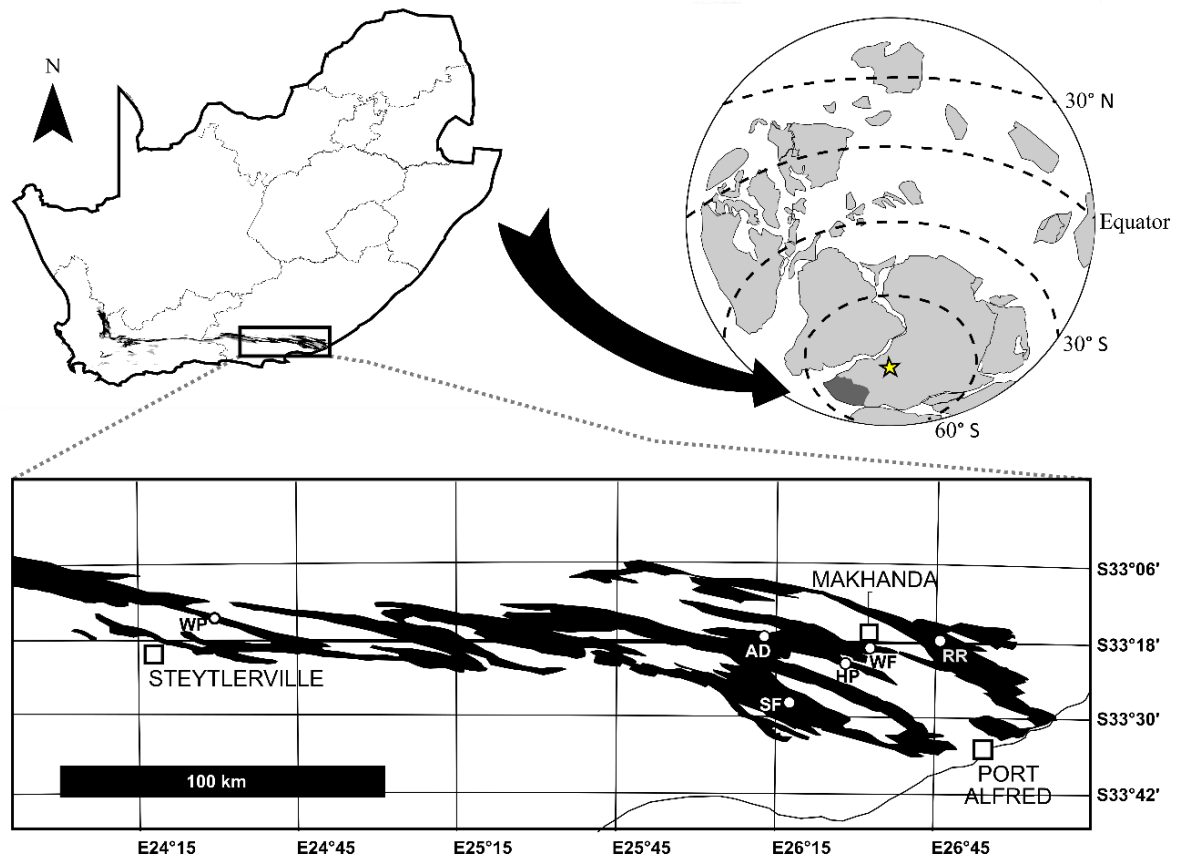


Figure 3.1: Map of the study area (bottom) showing its location in South Africa (top left) and palaeogeographic reconstruction for the Late Devonian (top right) after Torsvik and Cocks (2004), with the south pole indicated by a star. The study area is projected at approximately 70° South. Localities studied are Waaipoort (WP), Alicedale (AD), Seven Fountains (SF), Howison's Poort (HP), Waterloo Farm (WF), Grey Dam (at the same location as WF on map) and Rabbit Ridge (RR). The distribution of the Witpoort Formation is shaded black, based on data provided by the Council for Geoscience.

3.1.2. Trace fossils

Ichnotaxonomic descriptions are derived from trace fossils located in situ in each of the stratigraphic sections and in museum collections. A collection of 183 trace fossil-bearing rock slabs from the study area was accessioned into the catalogue of the Albany Museum, repositied at 87 Beaufort Street, Makhanda (Appendix 3). This material was collected by the writer and Dr Gess during fieldwork since 2015 and includes Witpoort Formation specimens from Coombs Hill (n=31), Rabbit Ridge (n=22), and Howison's Poort (n=4) as well as prior collections of Dr Gess from Waterloo Farm (n=77) and 'near Firglen turn-off' (n=3), and

Weltevrede Formation material from Howison's Poort (n=11). The provenance of each specimen includes the locality, stratigraphic horizon, collector's name and date of collection. Stratigraphic horizons for material described in Chapter 6 from Rabbit Ridge, Howison's Poort and Waterloo Farm refer to log diagrams presented in this work. Those from Coombs Hill refer to the stratigraphic logs presented in Harris *et al.* (2021a). The locality 'near Firglen turn-off' (33°19'12"S 26°22'2"E), approximately 15 km west of Makhanda, was investigated by Dr Gess during the excavation of a small farm dam in 1992. The locality was not revisited during this study. Satellite imagery and the Grahamstown 1:250 000 sheet (Council for Geoscience, 1995, sheet 3326) corroborate Dr Gess' assessment that the cutting is situated in the lowermost Witpoort Formation.

3.2. Methods:

3.2.1. Stratigraphy

Stratigraphic logs were measured using a Jacob staff and Abney level, or alternatively, where the dip of strata exceeded about 45°, exceeding the capability of the Abney level, a metre ruler was used. The logs were hand drawn in the field at a scale of 1.5 cm to a metre. These columns were redrawn digitally using Inkscape 1.0.1, with reference to field notes and photographs (Appendices 1B–1K). Structural tilt (dip angle and dip direction) of strata was measured using a Breithaupt geological compass with declination adjusted to -28°. Geological maps were drawn at approximately 1:10 000 scale with reference to field observations, existing 1:250 000 geological maps for the study area and satellite imagery (Google Earth). A Nikon Laser 550AS rangefinder was used to record distance measurements and slope angles. For calibration of thicknesses arrived at by manual logging, and to calculate the true stratigraphic thicknesses of intervals where strata were poorly exposed, the following formula was used (after Addie 1968):

$$TT = L \sin(AD \pm S) \cdot \frac{\cos(TD)}{\cos(AD)}$$

where TT is the true stratigraphic thickness

AD is the apparent dip of strata

TD is the true tip angle of strata

L is the length of traverse

S is the slope angle

3.2.2. *Sedimentary facies analysis*

The sedimentary facies analysis performed in Chapter 5 follows principles outlined by Dalrymple (2010). A sedimentary facies is thereby defined as a body of rock characterised by a particular combination of lithology, and physical and biological structures that bestow an aspect ("facies") that is different from the bodies of rock above, below and laterally adjacent. The characteristics used to define facies are generally those that have genetic significance. Accordingly, sedimentary facies do not necessarily imply a particular depositional environment, but reflect the conditions and processes responsible for their deposition. The facies then provide the building blocks for recognition of facies assemblages, which may, according to conventional facies models, reflect stacking patterns and lateral relationships characteristic of depositional environments (Dalrymple 2010).

Each bed is coded according to a two-letter facies code scheme (Miall 1977, 2010; 2014; Colombero *et al.* 2013), where a capital letter represents the dominant grain size of a lithofacies, and this is followed by a lower-case letter representing the characteristic sedimentary structure or texture (Miall 2010). According to the conventional grain-size code, sandstone facies are given an 'S', mudrock are given 'F', and grain sizes larger than sand are given 'G' for gravel. Twelve sedimentary facies are recognised in the study area (Table 5-1), including seven facies from Miall (2006), though five facies do not occur within Miall's scheme. Those not found in Miall's scheme include intraformational conglomerate (Gi),

extraformational conglomerate (Ge), undulatory sandstone and mudstone (Su), rhythmically-laminated sandstone with mud drapes (Sc), and soft-sediment deformation structures (SSDS).

Bedform geometry is one of the more definitive properties that separate facies, because this is the most detectable characteristic at outcrop and carries genetic significance (Lazar *et al.* 2022). Bedform geometry describes the thickness, lateral extent, nature of the bounding surfaces and internal stratification of the bedforms. Description of stratification follows the terminology formalised by Campbell (1967, p. 8) in which;

A lamina is the smallest megascopic stratum; a laminaset consists of a group of conformable laminae that form a distinctive structure within a bed; a bed is the stratum that reveals the principal rock layering, and a bedset consists of a number of superposed, similar beds.

This set of terminology contrasts with the more widely applied terminology of McKee and Weir (1953), in which ‘bed’ and ‘lamina’ are arbitrarily separated by having a greater or lesser thickness than 1 cm. In Campbell’s terminology, bed and bedset approximate the bedset and coset of McKee and Weir. For clarification, an example of a bed and a bedset as here used is shown in Figure 3.3. These units are described according to the quantitative terminology recommended by Campbell 1967, which are tabulated in Figure 3.2.

Rock types are classified according to mineralogy and grain size. Grain size is assessed with reference to a visual estimation chart based on the standard Wentworth scale (Fig. 3.2). Sandstones of the Witpoort Formation are fine to medium grained, composed of quartz, mica and clay with accessory quantities of heavy minerals and feldspar (Johnson 1991). The principal rock types encountered in the study area are intraclast conglomerate, quartz arenite, quartz wacke, siltstone and mudrock (laminated shale and mudstone) (Johnson 1991; Chabangu 2023). Shale is clearly separable from mudstone in having strongly developed lamination and exhibits a flaggy parting, whereas the latter is blocky. Rock samples from one

of the studied localities (WP) were collected by the writer and a postgraduate student, and were analysed petrographically (Chabangu 2023). Rock identifications are based on these results and prior petrographic work on rock samples from Coombs Hill in the study area (Harris 2017). According to the Pettijohn classification system (Pettijohn *et al.* 1973), quartz arenites are composed of more than 85% quartz, with less than 5% feldspar and/or lithic fragments. Quartz wacke contains between 15 and 75% clay-silt matrix, with the sand proportion comprising exclusively quartz and mica. Mudrock (>75% clay-silt) includes shale and mudstone, which are distinguished by the presence or absence of lamination respectively (Pettijohn *et al.* 1973; Schieber *et al.* 2010).

Rock colour is described qualitatively with reference to a colour chart (GeoExplore Store colour reference chart).

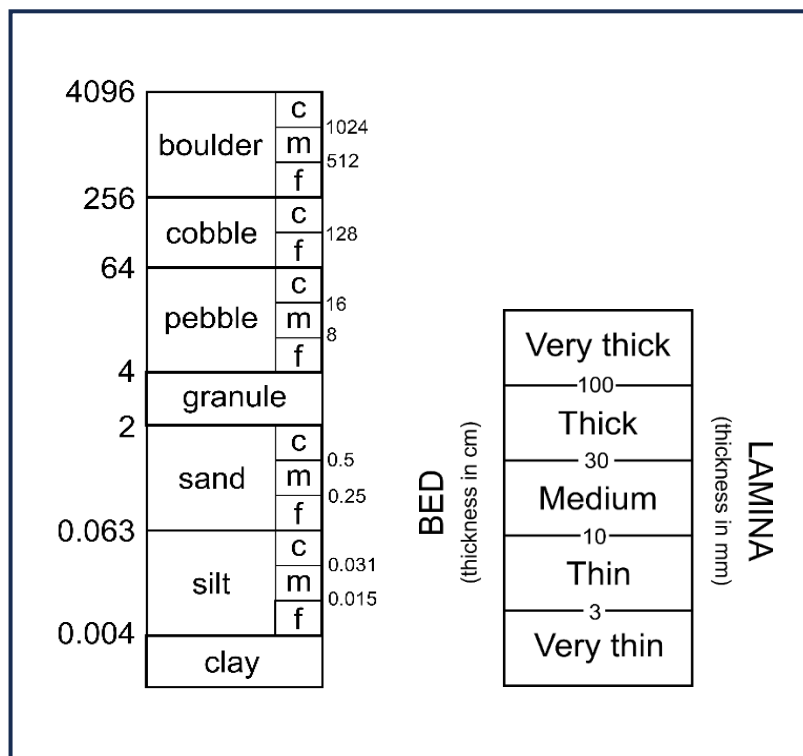


Figure 3.2: Left: The Wentworth Scale (grain size in mm). Right: Table of descriptive terms for “bed” and “lamina” after Campbell (1967).



Figure 3.3: Application of the terms Bed (B) and bedset (BS) to an example from the study area (Waaipoort). Beds within the bed set are weakly laminated, are separated by mud drapes, and thin upwards in the set. Scale increments are in cm.

3.2.3. Palaeocurrent analysis

Palaeoflow vectors (n=658) were measured in the field from *in situ* sedimentary structures. Prior to current analysis the vectors were corrected for tectonic tilt using a stereographic projection, according to standard procedure (Potter and Pettijohn 1977). Restoration is

achieved by interpreting, for each vector, the horizontal surface on which it formed ('bedding'), which then forms the axis of back-rotation. This bedding is restored to its original horizontal orientation by rotating first for fold axial plunge and then dip, and the flow vector is restored by the same rotation path. The data were restored in the program Stereonet (v. 11.0.9), and verified at random using a Schmidt stereonet projection. Palaeocurrent roses were generated in the same program with data bins set at 15°. The raw and rotated dataset is presented in Appendix 2.

In the study area, bedding measurements were taken from laterally extensive planar surfaces (flat surfaces of erosion or horizontally-laminated strata). As it was not possible to recognise such surfaces at every horizon, they were recorded where found. Vertically adjacent bedding surface measurements were averaged where variance in strike and/or dip was less than 10°.

Fold axial plunge was recorded as 3–4° in the Makhanda synclinorium (sections HP, GD and WF). Plunge was not corrected for in sections SF and RR. In the west of the study area, plunge was measured at between 9 and 12°. However, Booth (2002) recorded a plunge of 8° for the same locality. Given the larger sample set of bedding measurements from Booth's study, his result was used for the rotation. According to Potter and Pettijohn (1977), if axial plunge is less than 10° compensation for plunge is unnecessary. As a test, data from one of the sections (WP 2, plunge = 8° ESE), that with the steepest fold axial plunge recorded in the study area, was rotated with and without compensating for plunge (Fig. 3.4, A and B). These corrections had a small impact on the resulting palaeocurrent roses, altering the vector means by 5°, in accord with Potter and Pettijohn's evaluation.

Palaeoflow vectors were measured from directional indicators; cross-strata foresets, rib and furrow structures, trough axis orientations (cf. Michelson and Dott 1973), clast imbrications and scour tails, and asymmetrical ripple crests. Where no bedding surface

measurement of planar cross-strata could be attained, foreset dip was reconstructed on the stereonet using two apparent dip measurements, where the true dip and strike is defined by the great circle intersecting two lines of apparent dip on the same plane (Lisle and Leyshon 2004).

Bidirectional trend indicators, such as parting lineation, symmetrical ripple crestlines, plant fossil alignments and gutter casts are reported in the facies analysis (Table 5-1), but were excluded from the other analyses. Cross-strata that resolved at below 5° from horizontal were removed from the dataset.

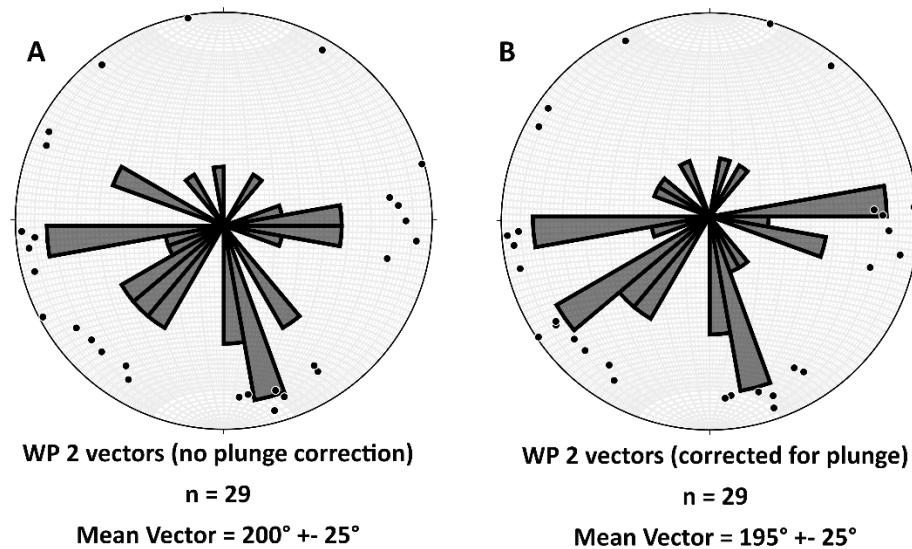


Figure 3.4: Flow vectors for Waaipoort, (section 2), without fold axial plunge correction (A), and with plunge correction (B). Black dots each represent a restored flow vector. Fold axial plunge for this section is 8° ESE (after Booth 2002). Error margins to the mean vector record one standard error.

3.2.4. Ichnology

Trace fossils were studied to better understand and interpret sedimentary processes, and for palaeo-environmental comparisons with other records. Accurate description and identification are important for this purpose. Trace fossils that could not be confidently assigned at the ichnogenus or ichnospecies level were placed in open taxonomy (Chapter 6, *incertae sedis*). Diagnosis of new ichnotaxa should be made with care because unwarranted

naming of trace fossils muddies the ichnotaxonomic literature and hinders palaeoenvironmental comparisons (Bertling *et al.* 2006) and it is considered beyond the scope of this study. The International Code of Zoological Nomenclature defines a trace fossil as the fossilised work of an organism (Bertling *et al.* 2006). Ichnotaxa are identified based on a morphological, rather than ethological, classification system (Mángano *et al.* 2002b; Bertling *et al.* 2006). Morphological traits that reflect the size or identity of the tracemaker, the substrate or the environment of deposition do not constitute ichnotaxonomic characters or ‘ichnotaxobases’ (Bertling *et al.* 2006).

Ichnofossils were measured directly, in the lab with a dial caliper, and in the field with a ruler, or indirectly from photographs using the digital software MB Ruler (v.5.4, 2022). Photographs were taken in the lab using a copy stand with low angled illumination. Photographic tone and contrast were altered using the program Adobe Photoshop CC 2017.

Degree of bioturbation is estimated with reference to the Bioturbation Index of Taylor and Goldring (1993), which was based on Reineck (1963). This index uses visual estimation to quantify the bioturbation intensity from BI 0 (no bioturbation) to BI 6 (complete bioturbation). This method has become standard usage by ichnologists and sedimentologists, enabling comparisons with established facies models (e.g. Buatois and Mángano 2011; Gingras *et al.* 2012; Dalrymple *et al.* 2015).

A bedding plane bioturbation index (BPBI) was developed by Miller and Smail (1997). It is a semiquantitative method describing the intensity of bioturbation on bedding surfaces using visual estimation. This method is less widely applied than the BI method, but serves for quantifying bioturbation that is limited to bedding surfaces and sometimes not visible in profile (Buatois *et al.* 1997b; Shchepetkina *et al.* 2019).

CHAPTER 4 – STRATIGRAPHY

4.1. Introduction

According to section 2.2 of the literature review above, lithostratigraphic units of the Witteberg Group have remarkable lateral persistence across the Cape Fold Belt (except where removed by Permo-Carboniferous subglacial erosion westwards and northwards). This, however, did not appear to be the case for the Rooirand, Perdepoort and Skitterykloof Members of the Witpoort Formation, which are only recognised westward of 25° E in the Cape Fold Belt (Thamm and Johnson 2006) and are mapped as far east as Kirkwood in the Eastern Cape (Sheet 3324, Council for Geoscience 1991). According to early work and subsequent reviews, the Witpoort Formation thickens from 300–400 m in the Western Cape to 850 m in the region of Makhanda (Johnson 1976; Theron 1994; Johnson *et al.* 2006). A more recent stratigraphic study in this eastern sector argued against this apparent thickening (Hiller and Taylor 1992). Moreover, structural geologists have suggested that increasing apparent thickness of Cape Supergroup units eastward was related to the greater degree of shortening and prevalence of thrust-related duplexing, and called for better integration between stratigraphic and structural studies (Shone and Booth 2005; Booth 2011).

In the eastern sector, Witpoort Formation stratigraphy is complicated by intensive structural deformation. Booth *et al.* (1999) documented the effects of asymmetrical folding, thrust faulting and normal faulting at Port Alfred to the southeast of Makhanda, demonstrating thrust-stacking of beds to be a particular hinderance to Witteberg Group lithostratigraphy. Based on the correlation of fossiliferous mudstone marker beds and the reversal of younging direction of strata across a tight, overturned syncline, Harris *et al.* (2021a) documented the duplication of an 80 m thick stratigraphic section at Coombs Hill to the east of Makhanda. Thickening of the formation in the vicinity of Coombs Hill inhibits the placement of its fossil-bearing horizons in the Witpoort Formation stratigraphic column,

precluding age comparisons for the site with Waterloo Farm (Harris *et al.* 2021a), the stratigraphic horizon of which is also not equivocal. Hiller and Taylor (1992) placed the Waterloo Farm black shale horizon in about the middle of an undifferentiated 350 m thick Witpoort Formation sandstone, while Gess (2011) considered the Waterloo shale to fall in the uppermost Witpoort Formation. New evidence from a systematic log of overlying strata will have implications toward resolving the disagreement. The fossiliferous horizons of Makhandia, including known sites at Waterloo Farm and Howison's Poort and a recently discovered locality at Rabbit Ridge are assembled in a composite stratigraphic framework, which hypothesises their positions in the column. by comparison to a reference section 200 km to the west.

The pass at Waaipoort exposes a full cross-section of the Witpoort Formation, which has been incorporated in prior stratigraphic (Johnson *et al.* 2006), petrographic (Olivier *et al.* 2011) and structural studies (Booth 2002). The sole published stratigraphic log showing the whole formation, including its upper and lower contacts, as well as the three members, is recorded from the site (Johnson *et al.* 2006). A re-investigation of this classic exposures provides a critical dataset for understanding the sedimentary and structural characteristics of the formation as a whole.

Based on the relationship of surface topography to underlying geological structure, some indication of Witpoort Formation stratigraphy can be formed. The Witpoort Formation is more competent than overlying and underlying formations and thus gives rise to prominent topographic highs in the study area. This topography is strongly influenced by structural folding of the strata, resulting in parallel WNW-ESE trending ridges that expose the formation. All strata in the studied sections are inclined parallel to the shortening direction, except for those occurring at the axes of fold structures, which, if the fold axis is open, may be sub-horizontal. Exposures of the lower Witpoort Formation tend to form 'inverted

topography' (Theron 1960, p.18), in which hills or ridges are underlain by synclinal structures. Adjacent anticline axes will form valleys due to uplift of the underlying Weltevrede Formation that is more easily eroded than the Witpoort. This is the case at lower Witpoort Formation localities at Howison's Poort (Fig. 4.1 B) and Rabbit Ridge. The normal condition, whereby an anticline forms a ridge, is observed wherever the upper Witpoort Formation is exposed in the study area, as found in sections measured at Waaipoort and Grey Dam (Fig 4.1 A, C).



Figure 4.1: Normal (A, C) and inverted (B) topography, illustrating the relationship of geomorphology to underlying geological structure and stratigraphy in the study area. A) An anticlinal ridge exposing the upper strata of the Witpoort Formation at Waaipoort. B) A ridge forming the southern limb of an open syncline exposing the lowermost strata of the Witpoort Formation at Howison's Poort. The syncline fold axis is approximately 300 m north (to left) of the Weltevrede–Witpoort Formation contact exposed in the treeline to the right of the photograph. C) The National Highway intersects the ridge defined by an open anticline exposing the uppermost strata of the Witpoort Formation at Grey Dam. The overlying Kweekvlei Formation is exposed in the syncline that defines the valley to the left of the photograph. The anticline axis intersects the ridge at the far right of the photograph, as mapped in Figure 4.8A.

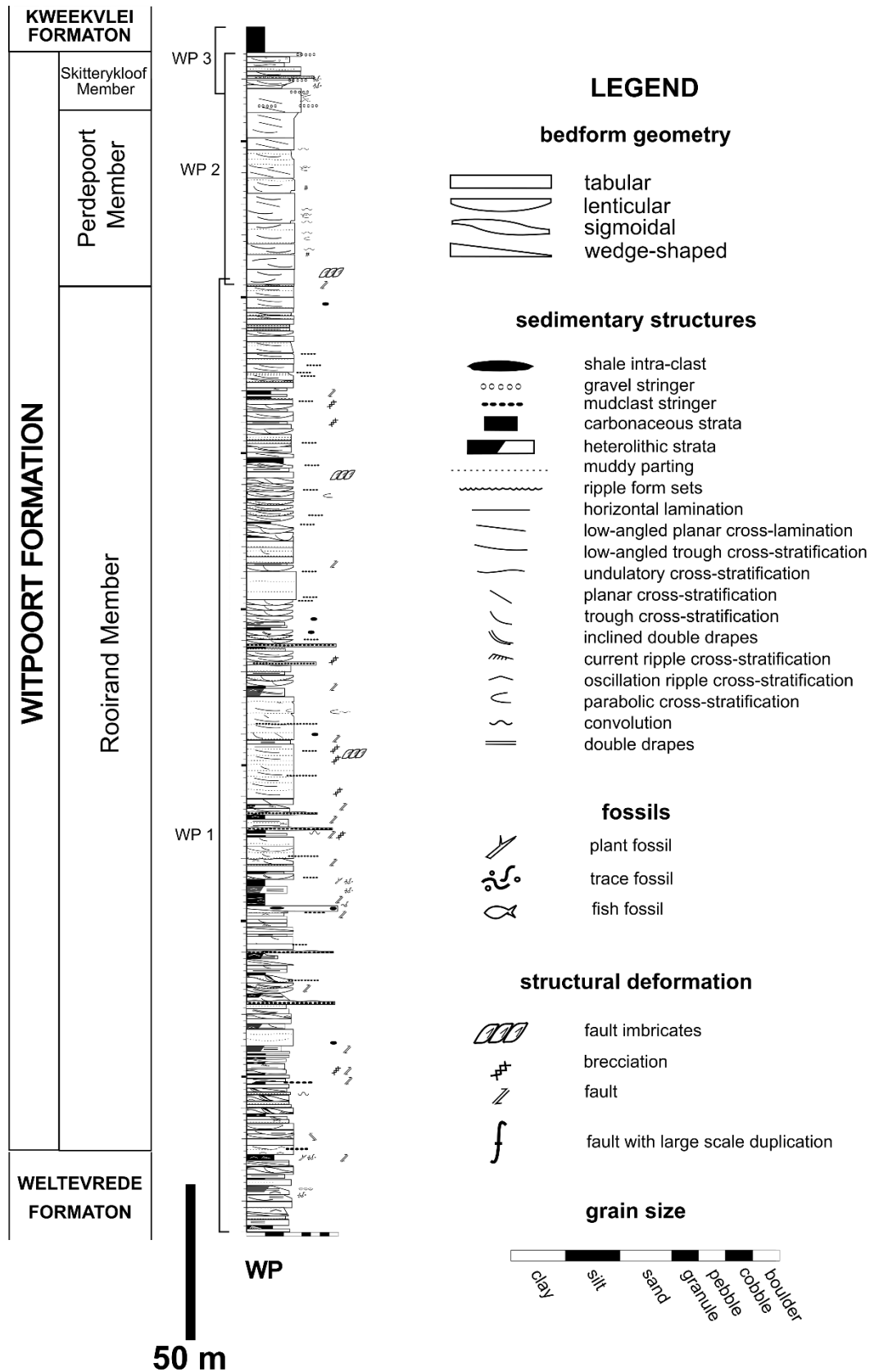


Figure 4.2: Composite stratigraphic log section for Waaipoort (WP 1–3)

4.2. Stratigraphy and structural geology of sections

4.2.1. Waaipoort

Horizon: upper Weltevrede Formation to lowermost Kweekvlei Formation

Description– The studied section is situated 190 km west of Makhanda, where a 1 km long continuous outcrop through Waaipoort exposes the upper Weltevrede Formation, the Witpoort Formation and the Kweekvlei Formation (Figs 4.2 and 4.4A). The strata are recorded in road cuttings and natural exposures along the R329 regional road, as well as in the adjacent valley formed by the Andries Buys River. Structural deformation is evident throughout the section, in places affecting the stratigraphy. The continuity of the stratigraphic section is interrupted by a series of thrusts and folds that repeatedly duplicate the upper strata of the Witpoort Formation towards the northern perimeter of the pass (anticlines A1 and A2, Fig. 4.4C). The strata were measured as a composite section (WP 1–3) to compensate for this deformation.

WP 1 is a continuous 307 m thick section through the upper strata of the Weltevrede Formation (25 m), the Rooirand Member (278 m) and lowermost Perdepoort Member, whereupon the section is interrupted with large scale faulting and a change in younging direction (Fig. 4.5A). Bedding in WP 1 is subvertical, dipping between 60°–80° NNE, and striking from 278°–294°. Small scale and medium scale faults with cm to m of displacement (Fig. 4.6A–C), slickensides and localised brecciation of strata affect the section throughout, increasing in frequency towards the large-scale thrust and fold axis that truncates the section. These small deformations often affect the superposition of strata, particularly around sandstone–mudrock contacts, and in places with duplication and/or fragmentation of beds (Fig. 4.6A and B).

WP 2 includes the strata of the Perdepoort Member (63 m thick) to the apparent top of the Skitterykloof Member (15 m thick), measured on the northern limb of A2 in the Andries

Buys River valley (eastern bank). The section begins at the axis of A2 where bedding dips at 30° NNE, progressively shallowing to about 10° NNE at the top of the Skitterykloof Member (Fig. 4.5B). Uppermost beds of the Skitterykloof Member are heavily deformed and faulted (Fig. 4.6D). Brecciation and weathering obscures their former contact with the Kweekvlei Formation, which has been erosionally removed from the top of WP 2. This contact is however observed in WP 3, where it is less deformed. WP 3 is measured on the sub-vertically dipping northern limb of A1 and records the uppermost 25 m of the Witpoort Formation, beginning at the first shale bed approaching the top of the Perdepoort Member, through the Skitterykloof Member to a conformable contact with the Kweekvlei Formation (Fig. 4.4C). The Skitterykloof Member, distinguished from the Perdepoort by the appearance of extraformational lithic clasts, is 20 m thick. Minor faulting affects this upper contact and another sandstone–shale contact lower in the section.

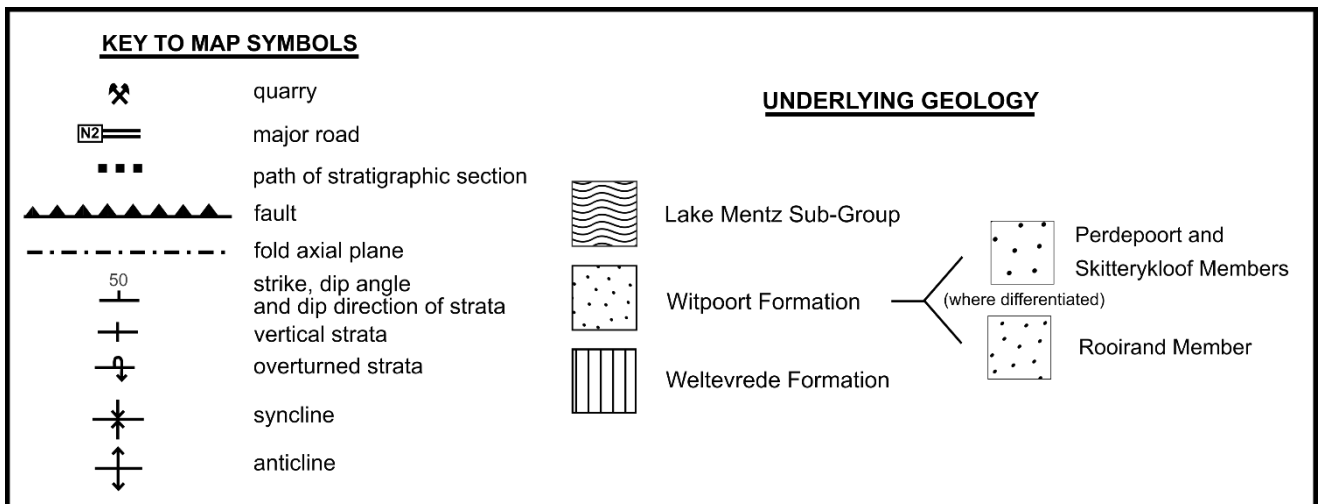


Figure 4.3: Symbolology for the maps presented in Chapter 4.

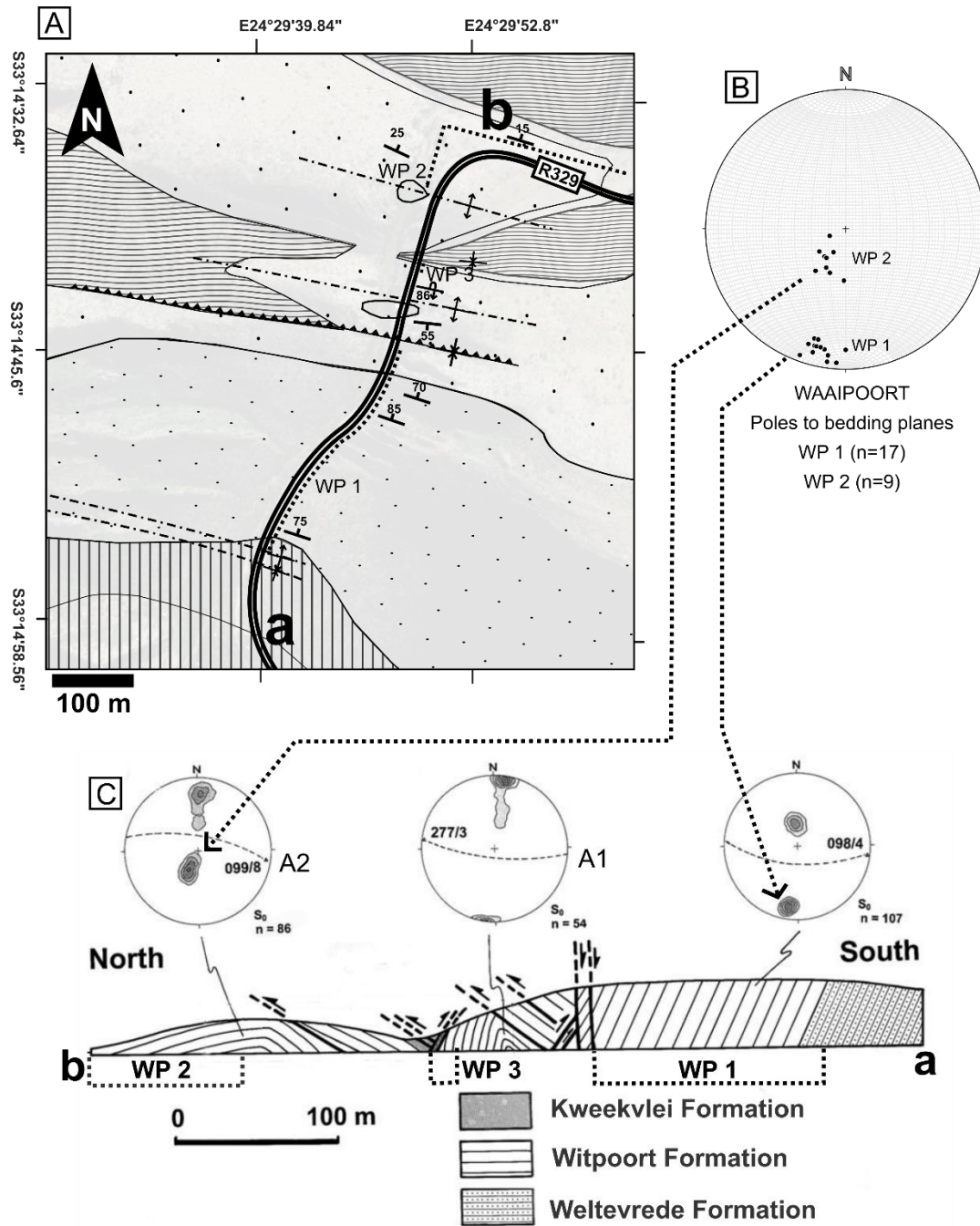


Figure 4.4: Location and geology of Waaipoort. A) Map of the area (for a key to symbols see Fig. 4.3). B) Stereonet projection of poles to bedding planes recorded in sections WP 1 and WP 2 (compare the data to the structural cross-section shown in C). C) Geological cross section (modified from Booth 2002) along transect a–b marked on the map in A. The stratigraphic sections WP 1, WP 2 and WP 3 are marked on Booth’s structural cross section.

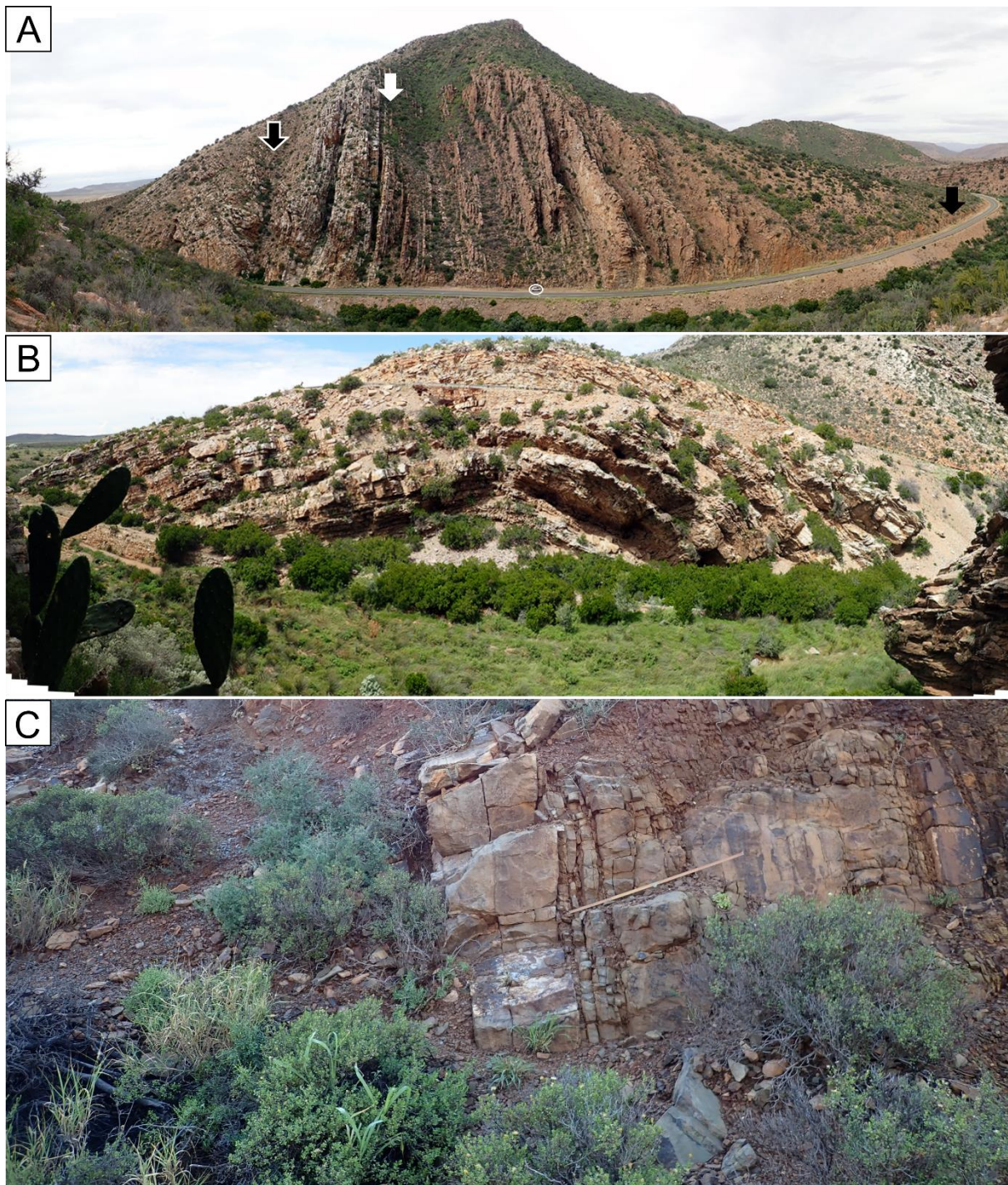


Figure 4.5: Studied exposures at Waaipoort. A) WP 1 is a continuous section from the uppermost Weltevrede Formation (black arrow at right) to the Rooirand-Perdepoort Members' contact (white arrow). A syncline/fault is indicated by a black arrow outlined in white. Anticline A1 is seen at far left of the exposure. Motor vehicle for scale is circled. B) WP 2 is recorded from the axis of anticline A2 along its northern limb to the top of the Witpoort Formation at far left of photograph. C) A sharp contact between the Witpoort Formation (Skitterykloof Member) and the Kweekvlei Formation shale is exposed in section WP 3. Younging direction is to left of photograph, ruler for scale is 1 m.

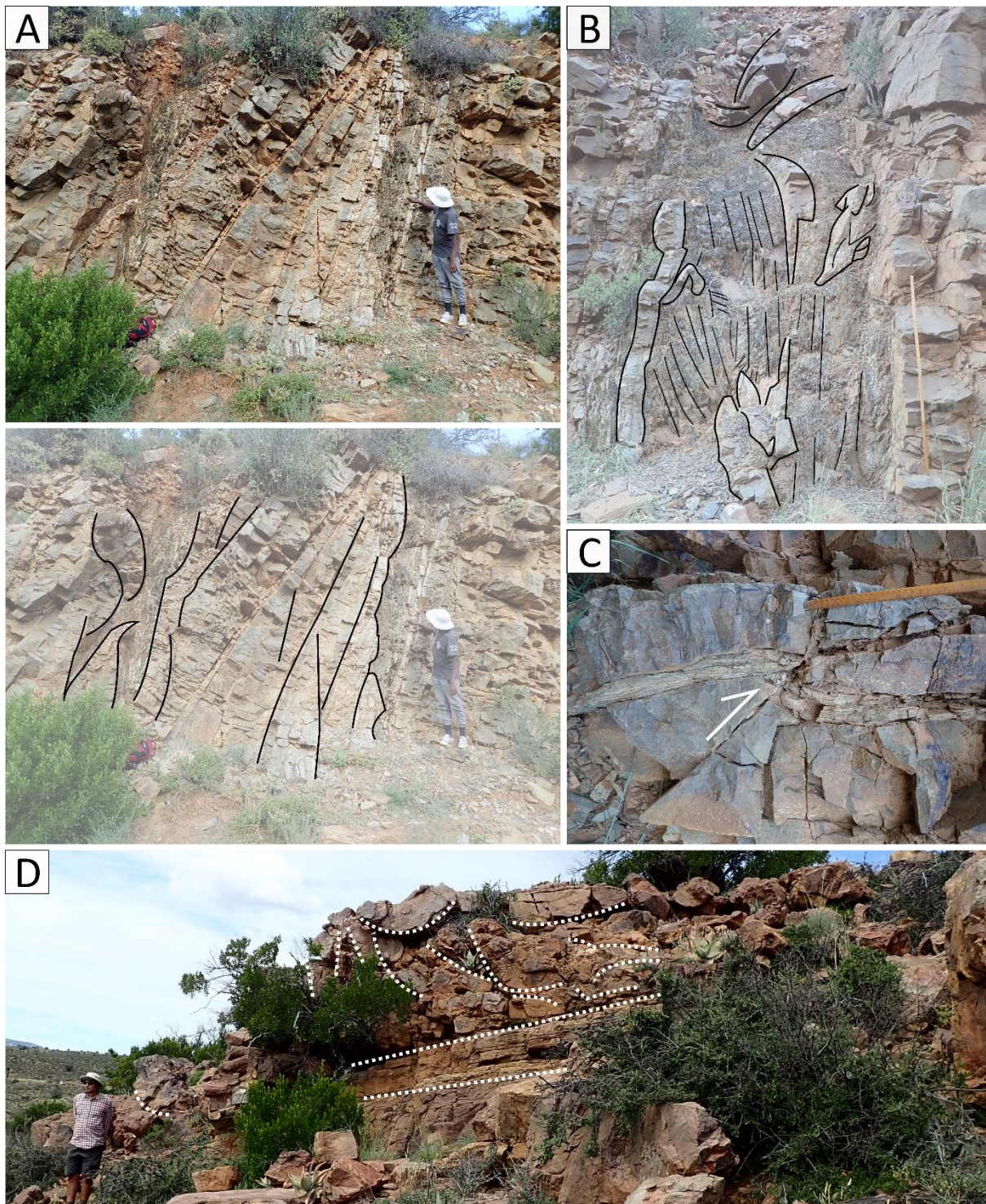


Figure 4.6: Structural fault damage at Waaipoort (A–C are from WP 1 and D is from WP 2). A) Photograph and line drawing (below) of a lower Witpoort Formation unit emphasising structurally deformed bed contacts (black lines). B) Ductile deformation of a sandstone interbed (thick lines) from a thick shale unit in which formerly horizontal lamination (thin lines) is disturbed. The sandstone bed (or beds?) is heavily fragmented. Ruler scale is 1 m long. C) A small scale fault with about 8 cm of displacement indicated by a white arrow. D) Deformation of the top of the Skitterykloof Member in WP 3. Bed contacts are emphasised with stippled lines.

Remarks– The structural geology of the Waaipoort pass was studied by Booth (2002). He recognised two oppositely verging anticlines in the section and recorded forethrust faults on the southern limbs of the folds (Fig 4.4C). The fault system that interrupts WP 1 was interpreted as a complicated system of thrust and backthrust faults that were later influenced by normal faulting associated with post-Palaeozoic extension in the basin (Booth 2002). Stratigraphic thicknesses here recorded are in close agreement with the section illustrated by Johnson *et al.* (2006), which was, presumably, also a composite section of the Waaipoort. The Rooirand Member is measured at 278 m thick, whereas Johnson *et al.* (2006, their Fig. 2) illustrate a thickness of approximately 280 m. The Perdepoort Member is 63 m thick in section WP 2 compared with 60 m shown by Johnson *et al.* (2006). The Skitterykloof Member is 15 m thick at the top of WP 2, and 20 m thick in WP 3. Though the section, particularly the Rooirand Member, appears to have been thickened by regular, small-scale thrust-related duplications it is, nonetheless, in the correct stratigraphic order and provides an important reference of sedimentary facies of the Witpoort Formation, its members and stratigraphic contacts to compare with the remainder of the study area in the eastern sector.

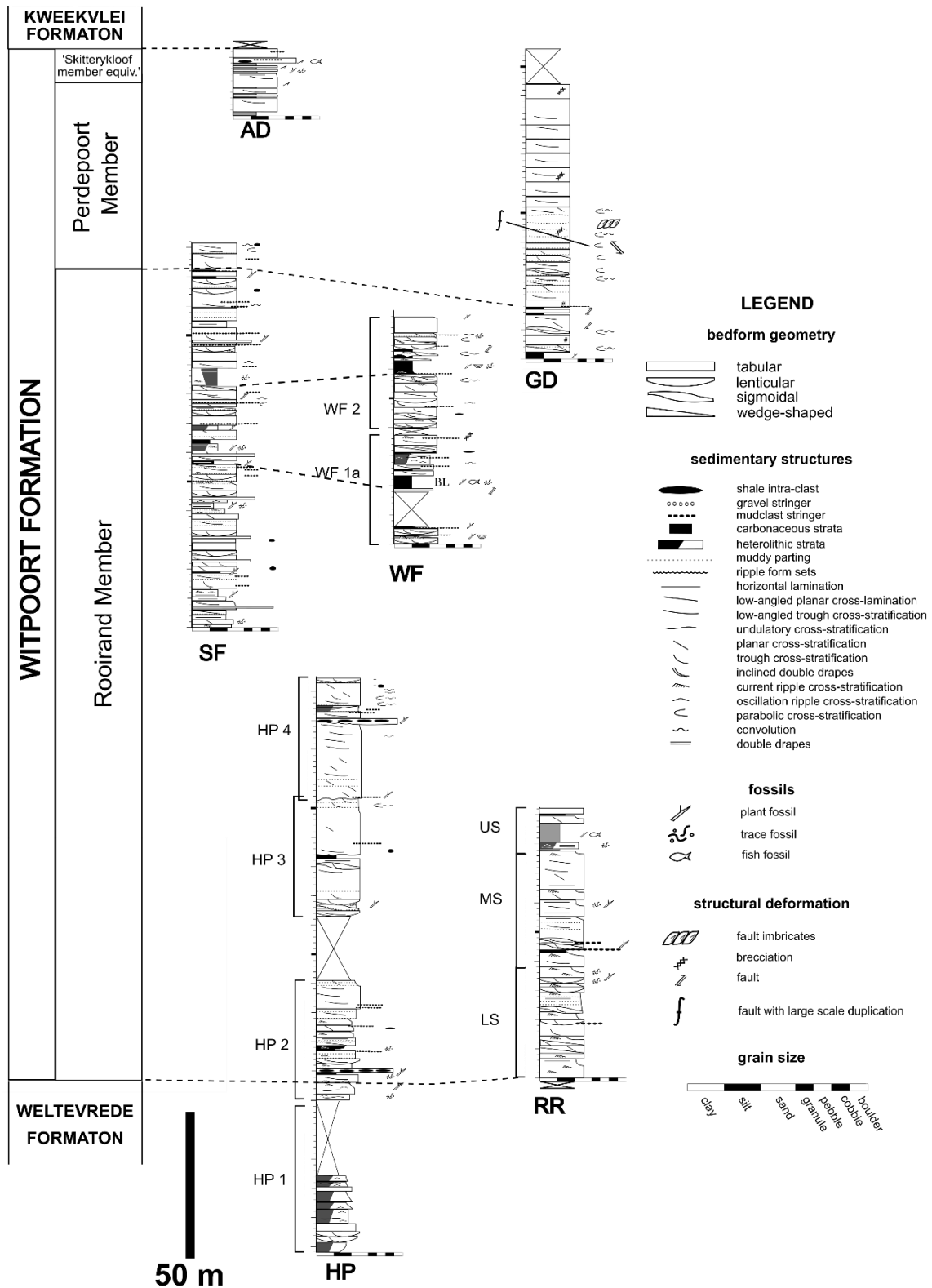


Figure 4.7: Composite stratigraphy for the eastern sector sections, Seven Fountains (SF), Alicedale (AD), Howison's Poort (HP), Grey Dam (GD), Waterloo Farm (WF) and Rabbit Ridge (RR).

4.2.2. Rabbit Ridge

Horizon: lower Witpoort Formation (Rooirand Member)

Description– Rabbit Ridge exposes a continuous composite section comprising three outcrops (*lower, middle and upper sections*) along a 1 km WNW transect adjacent to the N2 highway, approximately 20 km east of Makhanda (Fig. 4.8A). It records a stratigraphic thickness of 80 m of predominantly arenaceous strata underlying an unusually thick (~10 m), laterally extensive grey mudstone horizon (Figure 4.7; Appendix 1K). The strata in the section are structurally tilted, forming part of the SW dipping limb of a syncline. Bedding dip measurements are consistently between 60–80° (n=18; $\bar{x}$ =70°), striking at 290–320°, with no change in younging direction (Fig. 4.8B). The section begins at or near the base of the Witpoort Formation at the lowest stratigraphic exposure along this ridge. Strata in the section appear to be in stratigraphic order without disruption by folding or faulting. A zone of structural buckling to the northwest associated with brecciation and multiple changes in younging direction forms a negatively weathered anticlinorium that appears to have uplifted strata of the underlying Weltevrede Formation.

Remarks– The prominent Rabbit Ridge mudstone horizon was traced for 4.5 km along strike (see Chapter 7). It occurs at a similar stratigraphic level to conspicuous shale-rich horizons at Howison's Poort and Waaipoort, about 90 m above the base of the Witpoort Formation. The stratigraphy, sedimentology and palaeontology of this mudstone and underlying strata is given particular attention in Chapter 7.

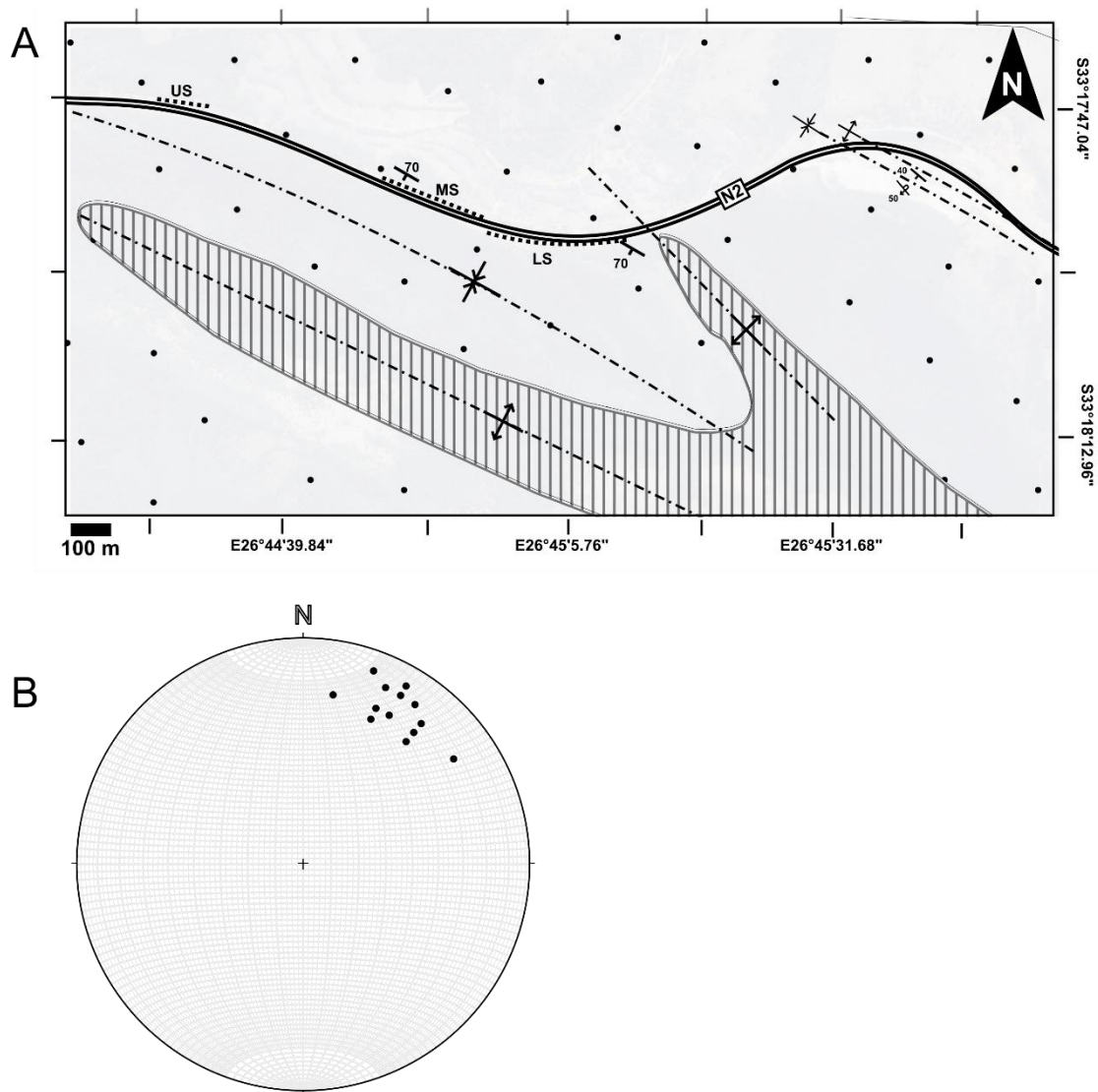


Figure 4.8: Geology and location of Rabbit Ridge: A) Map showing the locations of sections (Lower–LS; Middle–MS; Upper–US). The Rooirand and Perdepoort Members cannot be distinguished from satellite imagery in this area. B) Stereonet plot of poles to bedding planes from the studied sections.

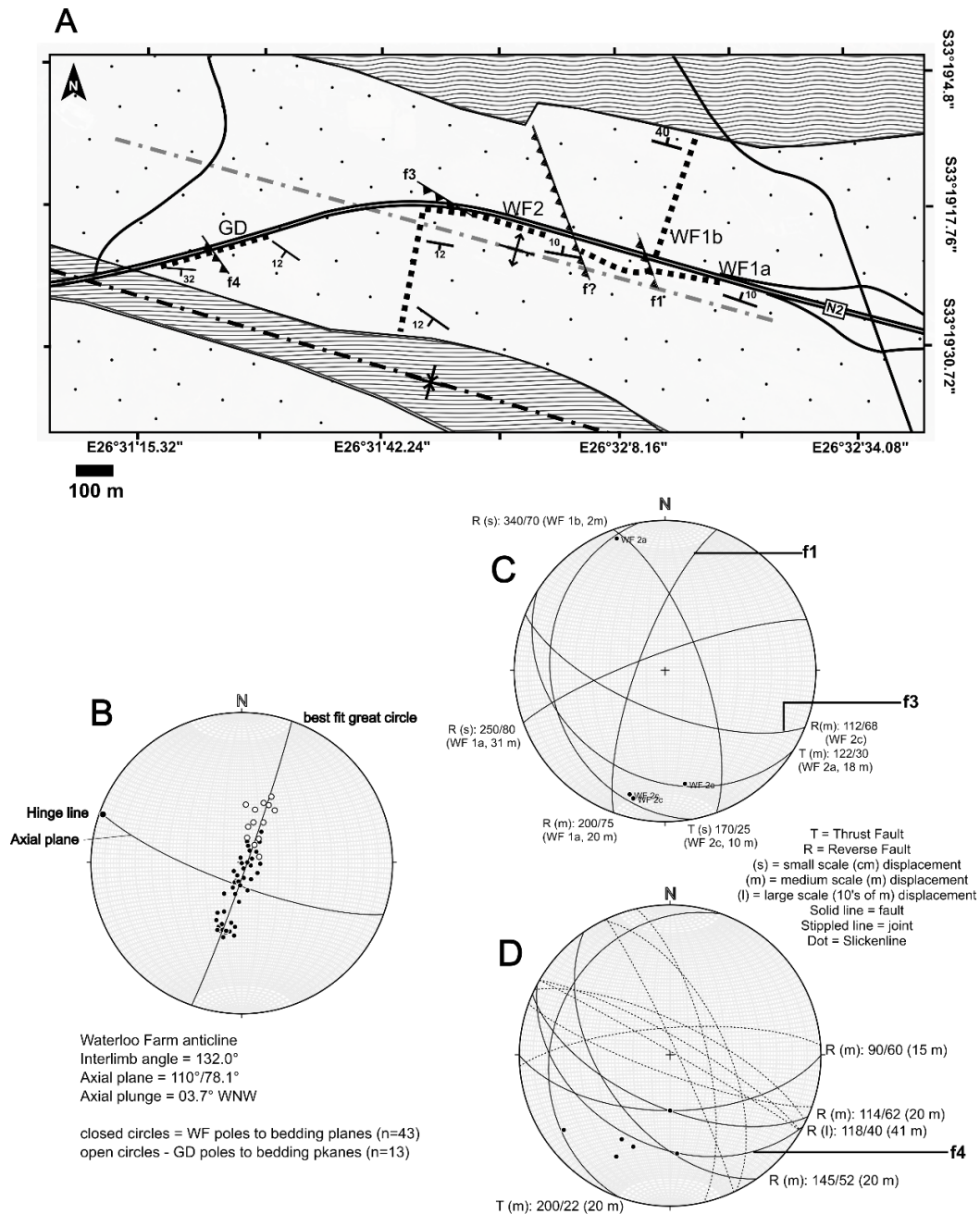


Figure 4.9: Geology and location of Waterloo Farm and Grey Dam sections. A) Map showing the locations of sections (Waterloo Farm WF 1a, 1b, 2) and Grey Dam (GD) (For a key to symbols see Figure 4.3). Though the Perdepoort Member is recognised in this area, it is not mapped separately from the Rooirand Member due to vegetation cover that obscures their contact in aerial imagery. B) Stereonet plot of poles to bedding planes from the sections. C) Faults from WF sections. D) Faults and joints from GD section.

4.2.3. Waterloo Farm

Horizon: upper Rooirand Member

Description– Three roadcut exposures along the N2 highway were measured, forming an 80 m thick composite section from the strata underlying the Waterloo Farm black shale lagerstätte to a horizon approximating the top of the Rooirand Member (Fig. 4.7). The overlying Perdepoort Member is recorded in the Grey Dam section, west of the Waterloo Farm sections.

The section is exposed in an open anticline (interlimb angle = 132°) shallowly plunging to WSW, with exposures of the overlying Lake Mentz Sub-Group to the north and south (Fig. 4.9A and B). Sections were measured in two directions on either limb of the anticline to compare the stratigraphic thickness to the top of the Witpoort Formation. The first (composite) section is measured westwards along north-facing road cutting exposures adjacent to the N2 highway and includes WF 1a, WF 2 and the Grey Dam section, providing an estimated 154 m from the fossiliferous horizon to the top of the formation (Fig. 4.7). The second section WF 1b is measured from the prominent black shale exposed in the south-facing cutting at Waterloo Farm, northwards towards Makhanda (Fig. 4.9A). Approximately 20 m of strata are recorded in the cutting overlying the shale, whereafter exposures on the hilltop above the cutting are intermittent. A green belt on the hilltop above this cutting may correspond to a 5 m thick shale recorded to the west in WF 2 on the opposing limb of the anticline, the Waterloo Farm ‘upper shale’. Northwards from this green belt the overlying strata, where exposed, are tabular, mature sandstones, much resembling those recorded in the Grey Dam section and are considered a lateral equivalent thereof. Section WF 1b gives a stratigraphic height of 170 m from the fossiliferous shale to the highest Witpoort Formation stratum on the northern limb of the anticline.



Figure 4.10: Structural faults in the Waterloo Farm section: A) Thrusting has emplaced underlying sandstones into a thick black shale bed (section WF 1a) several metres into the black shale, ruler is 1 m long. B) Enlargement of disturbed shale bedding from the top of the photograph in A. C) Thrust fault resulting in several metres of displacement of the thick shale/siltstone horizon from section WF 2. Bedding is approximately horizontal in the photo. D) Thrust stacking of sandstone overlying the Blue Layer (adjacent to section WF 1b). This structure occurs on the opposite side of the road to the thrust in A and shares a similar orientation of thrust propagation. Stippled lines indicate fault surfaces in A, C and D.

Exposures thus logged bear evidence of structural duplication. Several faults are recorded in the section including thrust faults, reverse faults and bedding parallel slickensides. The amount of bedding parallel slip associated with such slickensides can in places be constrained due to the presence of a straight walled, bedding-perpendicular clastic dyke (further described in Chapter 5) that crosscuts much of section WF 1. This dyke is displaced by bedding parallel slip at several horizons by between 30 cm and 1 m. Various decimetre faults are recognised with displacements measuring between 40 and 70 cm (Fig. 4.9C). They are not discussed further. Two prominent, metre-scale faults are recorded in the section with several metres of displacement, their orientations are shown in Figure 4.9C. One of them is a reverse fault that intersects a fossiliferous black shale horizon exposed in WF 1a (Fig. 4.10A). Intrusion of the underlying sandstone strata into the shale horizon has resulted in disturbance of the overlying shale (Fig. 4.10B). A similarly orientated fault occurs in the opposing cutting (Fig. 4.10D), which appears to be related to the same structure (mapped as f1 in Fig. 4.9A). In WF 2, a bedding oblique thrust fault intersects a prominent, 4–5 m thick shale horizon (f3 in Fig. 4.9A). This fault is observed on both sides of the highway. In the south facing cutting, the shale horizon is displaced by an estimated 6 m along the thrust plane (Fig. 4.10C). In the opposing cutting the fault exploits the contact between the shale and underlying sandstones. Given the metre-scale displacements along these observed faults, it is plausible that other similar, or greater structural displacements have impacted the section, which are not exposed. For example, at the top of WF 1 brecciation and ferruginisation of strata is observed. Overlying this, a vegetated interval of 3 m perpendicular to bedding could correspond to a fault of similar orientation to those described above, which is hypothesised based on satellite imagery ('f?' in Fig. 4.9A). Structural geology of the Waterloo Farm section is therefore complex and assessments of stratigraphic thickness in the overlying package are necessarily tentative.

Remarks– The black shale horizon forming the Waterloo Farm lagerstätte has been considered to be positioned stratigraphically in the uppermost Witpoort Formation (Gess and Whitfield 2020; Scholze and Gess 2021). The present analysis, however, places the fossil locality approximately 60 m below the Perdepoort Member. Quartzitic ridges on both limbs of the Waterloo Farm anticline overlying the site are recognised as the Perdepoort Member (see Grey Dam section below). These strata form the upper units of the Witpoort Formation as mapped by Goedhart (1:10 000 Map 3326 BC 6, Council for Geoscience 1996), rather than the Floriskraal Formation (Gess 2011, after Hiller and Taylor’s 1992 map). Comparable thicknesses (156 m vs 170 m) from the prominent black shale in section WF 1a and 1b measured to the highest exposed stratum on both limbs indicate that evident structural deformation in the sections has not disrupted the stratigraphy entirely, but may have influenced the thickness measurements locally. Vegetation cover along the ridges north and south of the locality have inhibited the distinction of the Rooirand and Perdepoort Members in satellite imagery and in resulting mapping. The more prominent faults in the Waterloo Farm sections are concentrated along sandstone-shale contacts. The fault-influenced black shale bed recorded above corresponds to the Blue Layer (BL, Gess and Prestianni 2022). It contains abundant and well-preserved plant fossils, but lacks vertebrate fossils. This horizon was recorded in section by Hiller and Taylor (1992, their Figure 11), according to the original 1985 roadcut, which was subsequently cut back twice, in 1999 and 2008, and partially consolidated with shotcrete. They show it to have been approximately 10 m thick. Hiller and Taylor correlated this shale with a layer then exposed to the east along strike, which contains fish fossils, and from which the bulk of palaeontological study on the locality is derived. The fish-bearing layer, the Main Fish Lens (MFL) (Scholze and Gess 2021) was only partially exposed during Hiller and Taylor’s study, giving a thickness of approximately 3–4 m. The layer thins and pinches out to the west (R.W. Gess pers. comm.; Fig. 4.11). The MFL is not

presently exposed, being covered with shotcrete. The shales were originally considered to represent the same stratigraphic horizon (Hiller and Taylor 1992), however, an angular, unconformable contact between the shale layers was observed by Dr Gess during the subsequent roadworks, who considers the layers to have been distinct. Immediately adjacent to the unconformity, a thin, inclined band of shale is yellowish in colour and exhibits shearing, resembling cross-stratification (R.W. Gess pers. comm.). The layers as they were exposed in 1990 are illustrated in Figure 4.11. Based on the reverse fault that intrudes the BL black shale as it is exposed today (Fig. 4.10A), it is likely that the deformed contact between these layers previously observed by Dr Gess is also fault-related. Differences in fossil biodiversity and taphonomy, and the colour and texture of the rock matrix between MFL and BL might indicate diachroneity or non-conformity between the layers. Apart from the lack of vertebrate remains in the latter, plant remains are larger and are more three-dimensionally preserved, and the matrix has a graphitic, bluish hue compared with MFL. Given the nonexposure of the contact, however, resolution of their exact relationship has not been achieved.

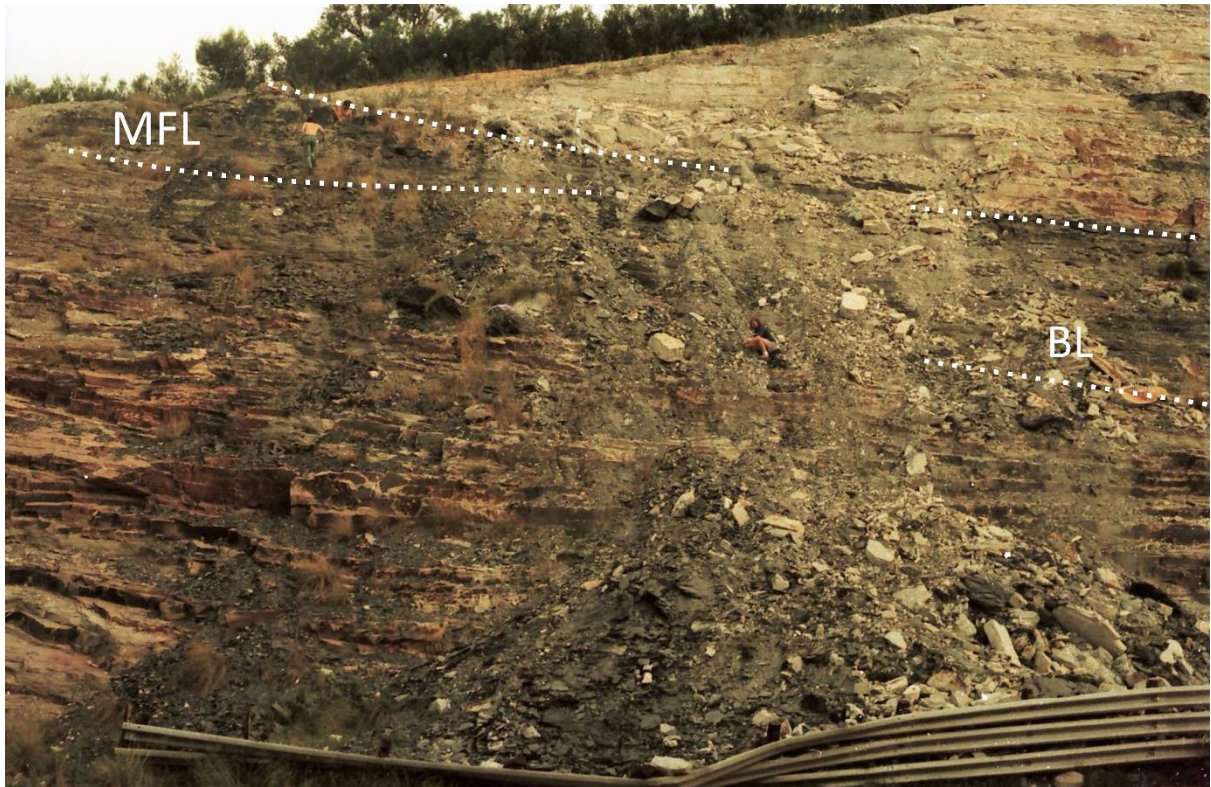


Figure 4.11: Photograph of the Waterloo Farm cutting, as it was exposed in 1990 (photograph provided by R. W. Gess). Stippled lines outline the two fossiliferous shale horizons, Main Fish Lens (MFL) and the Blue Layer (BL). The contact between these layers was obscured by scree in the photograph, and may have been influenced by faulting.

4.2.4. Grey Dam

Horizon: uppermost Rooirand Member and Perdepoort Member

Description– The Grey Dam section is a 92 m thick package logged in an approximately 250-metre-long road cutting adjacent to the N2 highway, 300 m west of the Waterloo Farm second cutting (WF 2) (Fig. 4.9A). It forms part of the south-southwest dipping limb of the Waterloo Farm anticline and bedding orientations on the stereonet plot along the axis-perpendicular great circle of that structure (Fig. 4.9B). This exposure presents a lithologically monotonous section dominated by extremely laterally persistent, tabular, cross-stratified sandstones. The top of the section is the uppermost exposure of the Witpoort Formation in the vicinity but the contact with the Kweekvlei Formation is not exposed. A sharp transition to shaly strata is suggested by a vegetated belt south of the Grey Dam section, which is mapped

as a syncline structure (Fig. 4.9A). Brown, immature siltstone and shale of the Kweekvlei Formation are exposed in a low roadcut about 500 m west of the Grey Dam cutting.

Numerous structures including small-scale thrust and reverse faults, a large-scale reverse fault (~tens of metres of displacement) and associated imbricate faults affect the section. Joint fields, bedding parallel slickensides (Fig. 4.12), localised brecciation and shearing of thin shale beds are prevalent throughout. Four out of five faults recorded in the section trend sub-parallel to the fold axis, and slickenlines on fault surfaces and bedding planes all show movement in the direction of structural shortening (Fig. 4.9D). Joint orientations show some consistency, trending approximately parallel to the fold axial plane (Fig. 4.9D). Several thin horizons (5–30 cm) of dark grey mudrock recorded in the section have undergone intense shearing, where the primary lamination of the rock is destroyed by phyllitisation.

A large-scale reverse fault (f4 in Fig. 4.9A) with displacement on the order of tens of metres is illustrated and interpreted in Figure 4.13. Several metres of strata in the footwall are obliquely truncated, whereas those on the hanging wall are thrust parallel to the fault. An increase of stratal dip by about 15° is recorded on the hanging wall. Imbricate structures and slickensides populate several metres of strata overlying the structure. This structure appears to have significantly duplicated the strata, but there are no marker beds with which to quantify the degree of thickening. Elsewhere in the section, zones of concentrated jointing obscure the sedimentary facies, for up to several metres in two places.



Figure 4.12: Slickenside on the lower bounding surface of a bed in the Grey Dam section (scale increments in cm).

Remarks– The stratigraphic position near the top of the formation and the sedimentary facies recorded therein identify the Grey Dam section as the Perdepoort Member, which was also studied at Waaipoort and Seven Fountains (4.2.1 and 4.2.7). A pair of shale horizons, 50 and 40 cm thick, 15–17 m above the base of the section is here taken as the Rooirand Member’s contact with the Perdepoort Member. An almost identical shale couplet occurs at this stratigraphic boundary in the Seven Fountains section, about 26 km WSW of Grey Dam. Above this horizon, the member is 80 m thick at Grey Dam (though it may have undergone significant structural thickening). The top of the section is brecciated and deeply weathered, explaining the apparent absence of a ‘Skitterykloof Member equivalent’ there. About 500 m north of the section, in a road cutting bordering Beaufort Street, that member might be

exposed on the north dipping limb of the same anticline. A Skitterykloof equivalent was studied at the Alicedale section (4.2.6).

The Grey Dam section is evidently thickened by thrusting, which explains the elevated thickness of the Perdepoort Member compared with Waaipoort where it is $\frac{3}{4}$ as thick. In both sections the member is characterised by extremely mature, tabular sandstone packages and a lack of thick accumulations of argillaceous facies or transported mudclasts. Typical for the Perdepoort Member, soft sediment deformation, particularly parabolic ('recumbent') cross-stratification and convolution, occurs commonly, affecting at least 30 beds in this section (see Chapter 5.1.11). Thin stringers of sheared mudrock occur at intervals of about 10 m in the Grey Dam section, forming notches in the section that separate arenitic composite sets that persist laterally for hundreds of metres (Fig. 4.13). Similarly formed units at Waaipoort are separated by negatively weathered silt-rich wacke interbeds.

The base of the Grey Dam section is a 2 m thick shale with plant and trace fossils, the bottom of which is not exposed. This bed might extend from the highest prominent shale in the Waterloo Farm section (WF 2), 300 m eastwards along strike, although this correlation is unconfirmed due to the intervening valley below Grey Dam, and because faulting is prevalent in the area. As the stratigraphically highest metres-thick shale horizon in the Witpoort Formation, this correlation is tentatively proposed (Fig. 4.7).

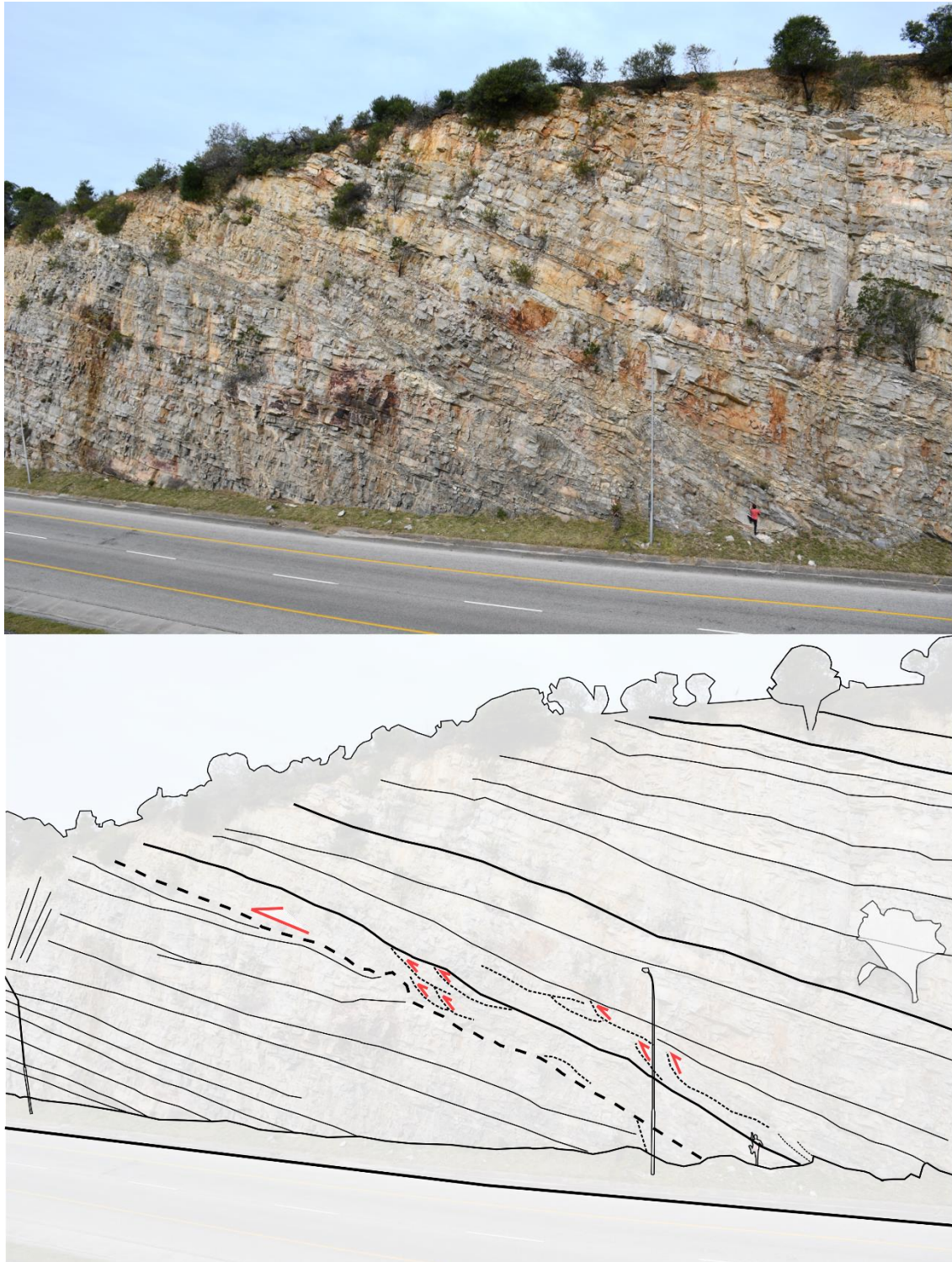


Figure 4.13: A large-scale reverse fault in the Grey Dam section with interpretive drawing below. The main fault plane is indicated with a stippled line. Note the tabular packages of quartz arenite above the fault that are typical for the Perdepoort Member. Person for scale at bottom right.

4.2.5. Howison's Poort

Horizon: upper Weltevrede Formation to middle Rooirand Member

Description– A 190 m thick succession of strata comprising the upper Weltevrede Formation (50 m) and the overlying lower Witpoort Formation (140 m) is logged at the locality. The sections include natural and roadcut exposures, forming a composite section that begins at the youngest prominent several metre thick sandstone unit in the upper Weltevrede Formation and terminates at the stratigraphically highest exposure in the vicinity at the axis of a syncline. The composite section includes four outcropping sections (HP 1–4; Fig. 4.14A; Appendix 1H) and intermediate stratigraphic gaps in which strata are not exposed. A sharp, conformable contact between the Weltevrede Formation and Witpoort Formation is exposed near the base of HP 2 (Fig. 4.15A). Localised shearing along this contact appears not to have disrupted the stratigraphic order. Discontinuities in the section, underlying and overlying HP 2, appear to be shale-rich intervals that are negatively weathered due to their relative softness.

The four sections are situated on the NNW dipping limb of an open syncline (interlimb angle 115°) shallowly plunging (3.7°) to the ESE (Fig. 4.14B). The dips of strata decrease gradually across the sections from about 35° at the base of HP 1 to approximately horizontal at the top of HP 4, which is situated at the fold axis. Apart from this fold, no structural duplications were recognised at the locality.

The few structures observed in the section include joints, quartz-filled tension gashes, slickenside surfaces and small-scale faults. Joints are approximately bedding perpendicular fractures, trending parallel, oblique and perpendicular to the fold axis (Fig. 4.14C). They occur locally at three horizons in the sections. At the top of section 2, fold axis-perpendicular megajoints cross cut tens of metres of strata, along which quartz crystals are locally developed. Slickenlines measured at a siltstone-sandstone contact in section 2 likewise trend perpendicular to the fold axis (Fig. 4.14B). The concentration of these structures proximal to

a 23 m stratigraphic gap between section HP 2 and HP 3 suggests structural damage as a plausible cause for the non-exposure of the strata. However, strata dips do not change across this stratigraphic gap. A concurrent increase in mud-draperies in the arenaceous strata at the top of the HP 2 indicates a typical upward-fining sequence, and it is hypothesised that a shale-rich interval provides a softer zone that has been exploited by the Palmiet River. The two explanations are not mutually exclusive, as structural damage is regularly concentrated around sandstone–shale contacts elsewhere in the study area.

Remarks—Howison's Poort is a historical Witpoort Formation locality. It was (informally) designated as one of three Witpoort Formation stratotypes by Johnson (1976) and was subsequently studied in detail by Hiller and Taylor (1992). There has been some confusion about the correct spelling of the name for this locality. This is due to an archaeological stone tool industry (the 'Howieson's Poort industry') that derives its name from a nearby rock shelter but uses an incorrect spelling for the poort's name (Smith 2008). The original spelling is here used.

Both Johnson (1976) and Hiller and Taylor (1992) recognised that the Weltevrede Fm–Witpoort Fm contact is exposed conformably at Howison's Poort. The latter study illustrated a stratigraphic log of approximately 100 m of the lower Witpoort Formation underlain by about 115 m of the Weltevrede Formation. During the present study it was recognised that intervals of non-exposure between the sections were not previously accounted for (Hiller and Taylor 1992).

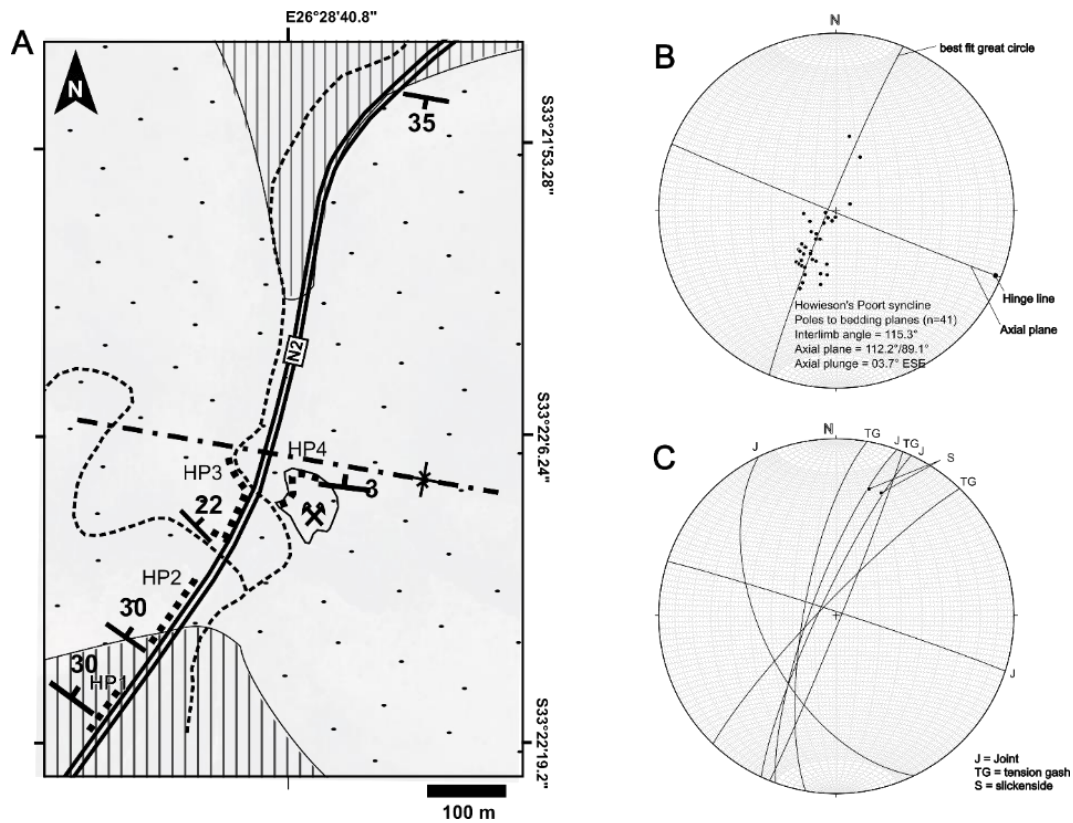


Figure 4.14: Howison's Poort geology. A) Map of Howison's Poort showing the paths of logged sections HP 1–4 (for a key to symbols see Fig. 4.3). B) Stereonet plot of poles to the bedding planes from the section with an interpretation of the syncline fold axis. C) Structures recorded in the sections.



Figure 4.15: Outcrop from HP and AD. A) Lowermost Witpoort Formation strata exposed at Howison's Poort (HP 2), the Weltevrede Fm–Witpoort Fm contact is shown with a dashed line. B) Vertically tilted uppermost Witpoort Formation strata exposed at Alicedale (younging to left). Bracketed area designates the logged section. The letter 'a' indicates an anticline structure duplicating the underlying strata, an asterisk indicates the conglomeratic horizon shown in C and D. The overlying Kweekvlei Formation shale (younging to the right) is exposed on the opposing limb of a syncline at far left of photograph. C) A 1.5 m thick conglomeratic unit overlies a 3 m thick shale-rich bed. Their contact (near bottom of photograph) is 5 m below the Witpoort–Kweekvlei Fm contact. Structural faulting and brecciation is evident in this exposure. D) Bedding plane view of weathered out, granule to fine cobble mudclasts (ruler is 50 cm²), photographed a few tens of metres to the right along strike from the approximate horizon of the ruler in C.

4.2.6. Alicedale

Horizon: uppermost Perdepoort Member and 'Skitterykloof Member equivalent'

Description– The lithological section near Alicedale was selected for study because it exposes a continuous 23 m thick section through the uppermost strata of the Witpoort Formation (Fig. 4.16B). The section is measured along the sub-vertically dipping northern limb of a tight anticline that duplicates underlying strata. The section begins at the first prominent shale horizon (approx. 1 m thick) above a several tens of metres thick succession of tabular, white-weathering quartz arenite, and terminates at the Witpoort Fm–Kweekvlei Fm contact. Kweekvlei shale bedrock is not exposed at the contact, however, typical fragments of orange-brown shale are found adjacent to the uppermost Witpoort Fm beds. Several metres of similar Kweekvlei Formation shale are well exposed in the axis of a syncline roughly 100 m north of the top of the section. The section is an arenitic, cross-stratified and structureless sandstone-dominated unit with mudrock interbeds. About 5 m below the uppermost Witpoort Formation stratum a 1.5 m thick conglomeratic bedset directly overlies a prominent, 3 m thick horizon of bioturbated, interbedded shale and sandstone (Fig. 4.15C). The conglomerate is very poorly sorted granule to boulder intraclasts embedded in a gritty coarse sand to granule sandstone matrix. A poorly preserved shark or acanthodian spine was found in the float adjacent to this conglomerate horizon (Fig. 5.2G).

The section is intensely deformed. Brecciation, local duplication of beds and shearing is profuse in the section, rendering the interpretation of sedimentary facies and flow vector orientation improbable, although the logged interval appears to be in stratigraphic order (Fig. 4.15B), whereas its underlying strata are multiply duplicated.

Remarks– The Alicedale section resembles the Skitterykloof Member recorded at Waaipoort (4.2.1) by its stratigraphic position underlying Kweekvlei Formation shale and overlying tens of metres of quartz arenite (the Perdepoort Member). With its shaly interbeds,

the section exemplifies an interfingering transition zone between these units.

Extraformational lithic gravel is not found in the Alicedale section and this separates it from the Skitterykloof Member, as recorded in two sections at Waaipoort (WP 2 and WP 3). The interval represents a lateral correlative of the Skitterykloof, a ‘Skitterykloof Member equivalent’ at the top of the Witpoort Formation (Fig. 4.7).

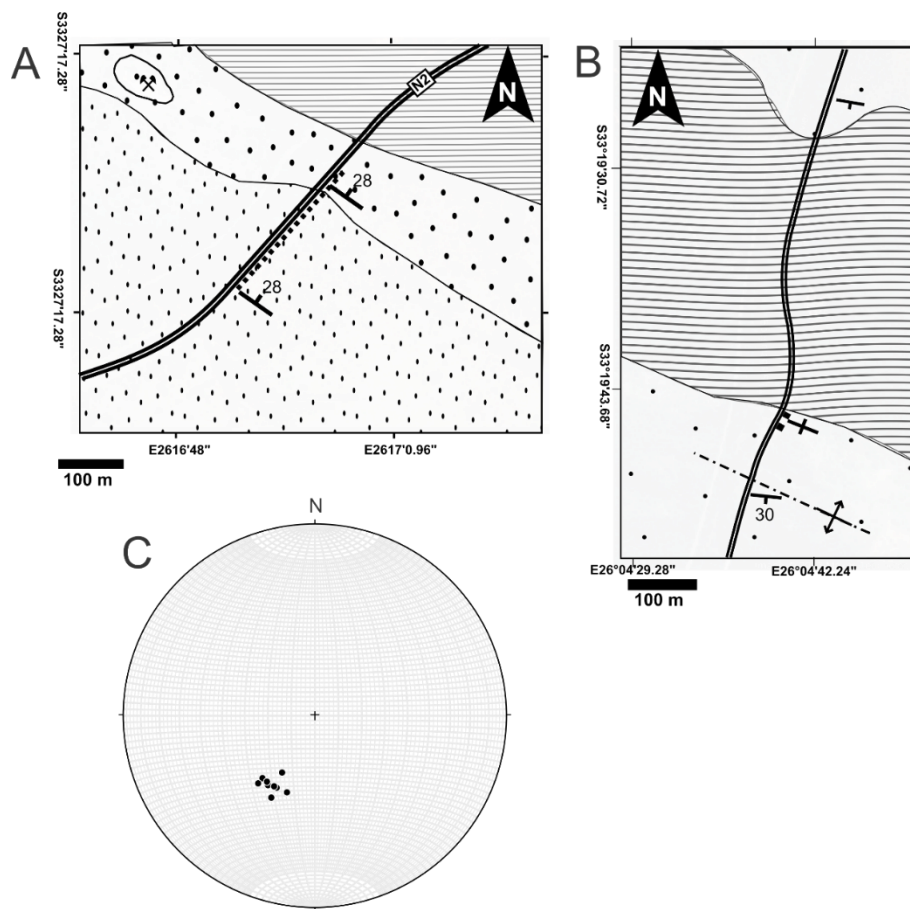


Figure 4.16: Locality maps for A) Seven Fountains, B) Alicedale, and C) Stereonet projection of poles to bedding planes from Seven Fountains (for a key to map symbols see Fig. 4.3).

4.2.7. Seven Fountains

Horizon: upper Rooirand Member to lowermost Perdepoort Member

Description– The Seven Fountains section is a formerly undescribed geological section that derives its name from an 18th Century farm that is situated 5 km northeast of the exposure. It records 125 m of northeast dipping upper Witpoort Formation strata, measured in

a continuous 250 m long road cutting adjacent to the N2 highway 26 km SE of Makhanda (Fig. 4.16A).

The Rooirand Member and Perdepoort Member are clearly distinguishable by a contrast in weathered rock colour (Fig. 4.17), as observed at Waaipoort. The red-brown weathering, upper Rooirand Member is a 115 m thick sandstone rich assemblage with mudrock at intervals, containing plant and trace fossils. It is overlain by the white, arenaceous Perdepoort Member, the lowest 10 m of which is exposed in section. The section is truncated at the top by a fault (Fig. 4.17). The top of the Witpoort Formation is inferred by a sharp sandstone–shale transition, 60 m NE of the section, the Witpoort–Kweekvlei Formation contact (according to the 1:250 000 Grahamstown Sheet, 3326, Council for Geoscience, 1985). These lithological boundaries are detectable in satellite imagery and are illustrated in Figure 4.16A. At the base of the section, underlying Witpoort Formation strata are covered by Quaternary alluvium.

Strata have consistent dips of about 35° to NE (Fig. 4.16A and C), which vary slightly in the vicinity of structural faults (28° – 40°), and strike between 290° and 310° (Fig. 4.16C). Two prominent faults are observed in the section. A reverse fault is recorded at 70 m above the base of the section, comprising a pair of parallel, sub-vertical N–S trending fault planes containing a brecciated zone. Adjacent strata show evidence of bedding parallel slip but appear to be in sequence. At the top of the section the dip of Perdepoort Member strata abruptly increases to vertical, and this is associated with brecciation and scree cover, whereupon the roadcut exposure terminates (Fig. 4.17).



Figure 4.17: Seven Fountains section outcrop. Top photo illustrates upper Rooirand Member strata seen from the base of the section. Bottom photo shows the contact between the Rooirand member (right) and Perdepoort Member (left), indicated by a stippled line. Younging direction is to the left. The top of section SF is at left of bottom photograph.

Remarks– The section is well-preserved and exposes a stratigraphic interval of the upper Rooirand Member that is lithologically comparable to that exposed in the Waterloo Farm sections. Two prominent mudrock horizons occur in the section at approximate horizons corresponding to the BL and upper shales at Waterloo Farm, which are hypothesised to

represent a proximal-distal profile. A metre-thick, conspicuously dark, carbonaceous horizon in the section occurs 56 m below the Perdepoort Member, almost precisely at the same stratigraphic level as inferred for the Waterloo Farm BL and MFL (Fig. 4.7). These argillaceous horizons differ in the Seven Fountains section in being more heterolithic compared to the more monolithic black shale horizons of Waterloo Farm, which have better preserved fossil plants as well as fish. Common large plant axes (up to 6 cm in diameter), associated with mud-pebble lag deposits in the quartzitic strata, are a feature of this stratigraphic interval at Seven Fountains and at Waterloo Farm, but were found nowhere else in the study area (Figs 5.23 and 5.30).

4.3. Stratigraphy of the Witpoort Formation in the Eastern Cape

4.3.1. Lower contact

The Weltevrede Formation–Witpoort Formation contact is sharp in the two places that it was studied. It is very clear at Howison's Poort, with an abrupt shift from mudrock- to sandstone-dominated deposits (Johnson's 1976 criterion). At Waaipoort both the Witpoort and the Weltevrede are more heterolithic, but the sharp shift is also recognised there. This lateral change can be related to basinal factors (i.e. local environment, variable subsidence or proximal-distal gradients) along a synchronous or diachronous contact. These possibilities will be elaborated upon based on the facies assemblages, sediment dispersal patterns and ichnology across the contact.

4.3.2. Rooirand Member

The Rooirand Member forms the bulk of the Witpoort Formation. The member, which is 278 m thick at Waaipoort, predominantly comprises quartz arenite and quartz wacke, with a subordinate proportion of shale occurring at intervals throughout.

In the Waaipoort and Seven Fountains sections, the Rooirand Member is visibly distinguishable from the Perdepoort Member by its reddish-brown colouration, which

contrasts markedly with the whiteish colour of the latter (Figs 4.5A, 4.17). The reddish colouration of the former is derived from mineral precipitation along fractures on the rock faces (Fig. 4.18). East of Seven Fountains colour differences are less marked. In all sections the Rooirand Member exhibits metre to tens-of-metre scale cyclicity with intraformational conglomerate, amalgamated tabular sandbodies, lenticular sandbodies and subordinate mudrock. These lithologies occur in repetitive upward-fining (muddier-upward) cycles.

In the eastern sector the Rooirand Member is exposed at Rabbit Ridge, Howison's Poort, Waterloo Farm and Seven Fountains, and was also previously studied at Coombs Hill (Harris *et al.* 2021a) and Woest Hill (Hiller and Taylor 1992) in the vicinity. The lower portion of the member is exposed at Howison's Poort (140 m) and at Rabbit Ridge (90 m). The upper portions of the member are exposed at Seven Fountains (115 m) and at Waterloo Farm (80 m). The combined 255 m composite section is slightly less in thickness than the 278 m thick Rooirand Member at Waaipoort. Parallels can be drawn with the section at Waaipoort. The muddier-upward cyclicity is most evident in the lower 80–100 m of the member and the upper ~120 m of the member, but is lost within a sandstone-dominated middle portion where stacked, tabular quartz arenite attains thicknesses of 50 m at Howison's Poort (HP 4) and 35 m at Waaipoort. Prominent mudstone and shale horizons occur at approximately 80–90 m above the base of the member at Waaipoort, Howison's Poort and Rabbit Ridge. Likewise, towards the top of the member, carbonaceous black sandstone and shale horizons occur at similar levels underlying the Perdepoort Member, as observed in sections at Waaipoort, Seven Fountains and Waterloo Farm. These apparent correlations across the study area suggest that development of mudrock may have been controlled at a basin-wide scale, and contradict the notion that the Witpoort Formation stratigraphy considerably thickens to the east. If the formation thickened eastwards due to increased syn-depositional subsidence, it should be reflected in the cyclicity and rate of sedimentation. Equal accommodation would be

reflected in correlatable facies changes in stratigraphic packages of similar thicknesses, as appears to be the case in the study area.



Figure 4.18: A thick bed of massive sandstone from the upper Rooirand Member at Waaipoort showing the oxide minerals that have precipitated along fractures in the rock face, giving the member its reddish-brown colour.

4.3.3. Perdepoort Member

The Perdepoort Member attains a measured thickness of approximately 56 m at Waaipoort (Fig. 4.2) and 80 m at Grey Dam (Fig. 4.7). In the Western Cape thicknesses of approximately 35 m (near Gamkapoort) and 55 m (near Willowmore) were previously

reported (Haughton *et al.* 1953, in Looock 1967). The Perdepoort Member therefore appears to thicken eastwards. However, the 80 m figure measured at Grey Dam appears to have been influenced by structural thickening and an uncertain amount of duplication.

The Perdepoort Member is a more petrographically mature sandstone unit than its overlying and underlying members (Olivier *et al.* 2011). It lacks the cyclical facies changes of the Rooirand Member in favour of lithologically monotonous cross-stratified sandstone with abundant soft-sediment deformation. The member is developed in extremely tabular packages that are separated by slight notches (Fig. 4.18). Fossils and trace fossils, prominent mudrock or intraclasts are not found in the member in the studied sections.

The member is comparable across the study area. At Waaipoort (WP 1) its lower contact is transitional, with a shift of red-brown to white that is highly visible at outcrop. This level also corresponds to the last shaly interbeds at the top of the Rooirand Member. Similarly, at Seven Fountains the contact marks a colour shift corresponding to the disappearance of shaly interbeds. At both Seven Fountains and Grey Dam the contact is marked by a conspicuous pair of medium to thick beds of laminated shale. The Perdepoort Member is transitionally overlain by the Skitterykloof Member and its correlative to the east.



Figure 4.19: The Perdepoort Member at Waaipoort. Note the extremely tabular packages separated by slight notches. The Skitterykloof Member is at the level of the top of the cliff. Lake Mentz Subgroup strata are exposed in the background, north of the section.

4.3.4. Skitterykloof Member

At Waaipoort the Skitterykloof Member is a 15–20 m thick succession of coarse-grained sandstone containing stringers of lithic clasts. The introduction of quartz extraclasts is gradational, often localised, and does not correspond to a sharp facies shift from the Perdepoort Member. Shaly and heterolithic interbeds in the member signify the most conspicuous facies change. In two sections at the locality (WP 2 and WP 3), the largest extraclasts, up to coarse pebbles, occur above a bioturbated siltstone bed about 10 m below the top of the member.

Extraformational lithic clasts are not apparent at the top of the Witpoort Formation in the eastern sector, but a similar succession of facies with the reintroduction of bioturbated

mudrock and intraformational conglomerate, marks this unit as a correlative interval here named the ‘Skitterykloof Member equivalent’ (i.e. Alicedale section). A couplet of bioturbated shale and intraclast conglomerate potentially represents a marker ‘event bed’ 5 m below the top of the formation in the section. The ‘Skitterykloof Member equivalent’ is not exposed at the top of the formation in the vicinity of Waterloo Farm where it may have undergone décollement from the more competent Perdepoort Member. Conglomeratic beds at the top of the Witpoort Formation succession in the region have been observed by R.W. Gess (pers. comm) on the farm Hilton (east of Makhandia) and at sites along Governor’s Kop eastward of Makhandia indicating a wider expression of this ‘Skitterykloof Member equivalent’. This unit, as a putative correlative of the Skitterykloof Member, is a priority for future investigation to understand its lateral variability and stratigraphic significance (see Chapter 8.6).

4.3.5. Upper contact

A sharp contact with the Kweekvlei Formation is inferred everywhere in the study area, but the contact surface was only seen at Waaipoort where it overlies the Skitterykloof Member with the sudden onset of continuous laminated shale deposits (Fig. 4.5C). At Alicedale it overlies a Skitterykloof Member equivalent where a similar, sharp lithological contact is indicated. Weathering and soil cover obscures the contact in the vicinity of Waterloo Farm, where the contact appears to have been exploited by structural décollement and negative weathering. The Kweekvlei Formation shale is tens of metres thick across the study area and in the wider Cape Basin (Browning and Reid 2016). Although its direct contact with the Witpoort Formation has few exposures in the basin (J.C. Looock, pers. comm., 2020), it can be clearly recognised from satellite imagery throughout the study area.

CHAPTER 5 – SEDIMENTARY FACIES ANALYSIS

5.1. Introduction

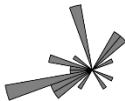
In the preceding chapter it was argued that the Rooirand Member, Perdepoort Member and Skitterykloof Member (or its equivalent unit) are present across the study area, which provided a stratigraphic framework for the provenance of the rocks and ichnofossils that are described in this and the forthcoming chapter respectively.

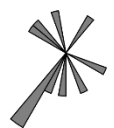
In 1992 the Hiller and Taylor hypothesis of a mesotidal, wave-dominated barrier island system for the Witpoort Formation's eastern sector was proposed. Then, based on decades of palaeontological research on fossil concentrations from a black shale horizon at Waterloo Farm, Gess and Whitfield (2020) interpreted the deposit as having been an estuary, with periodic salinity stratification and bottom water anoxia, rather than a stagnant lagoon as previously thought (Hiller and Taylor 1992). Sandstones hosting the black shale horizon are thought to be deposits of back-barrier tidal channels and washover flats (Hiller and Taylor 1992; Gess and Whitfield 2020).

Tidal processes, if important in the study area, should be reflected in the strata. Their resulting deposits should be evidenced by the sedimentary facies, sediment transport patterns and ichnology. Progress in coastal sedimentology theory since 1992 with facies models (Walker and James (eds) 1992) and subsequent refinements, that are reviewed in this chapter, are revolutionising one's ability to recognise the complex array of sub-environments in ancient paralic successions. Deposits of tidal inlets and tidal deltas, beaches and shorefaces would be recognisable according to stacking patterns that reflect their characteristics (Reinson 1992), which have not been adequately demonstrated before in the Witpoort Formation. This barrier island model, as well as open-marine deposition ascribed to the wider Witpoort Formation, will now be tested (Broquet 1992; Hiller and Taylor 1992; Cotter 2000; Shone and Booth 2005; Thamm and Johnson 2006; Johnson *et al.* 2006; Olivier *et al.* 2011;

Harris *et al.* 2021a). The aim of this chapter is to investigate the processes of deposition by describing sedimentary facies of the studied strata. The palaeoenvironmental interpretation will be based on vertical and lateral relationships of the facies, their flow vectors and palaeoecology, which form the bases of the Facies Assemblages that are the culmination of the present chapter.

Table 5-1 (next two pages): Sedimentary facies, their main characteristics and facies codes.
Abbreviations: L – lamina; B – bed; BS – bedset; F – fine-grained; M – medium-grained; C – coarse-grained; lbs – lower bounding surface

Code	Lithology and sedimentary structures	Thickness / lateral extent	Flow vectors	Process
Gi	Intraclast conglomerate Poorly sorted granule to fine boulder sized intraformational sand/mud clasts in a quartz wacke matrix, clasts deformed	B: thin to thick / a few metres-10's of m commonly localized scour fill	no data	Scour fill / Lag deposit
Ge	Extraformational conglomerate Variably sorted granule to coarse cobble-sized lithic quartz pebble stringers in quartz arenite or wacke matrix	B: one to several clasts thick / lateral extent uncertain	no data	Lag deposit and/or deflated surface
St	Trough cross-stratified sandstone F-M quartz arenite, curved upwards concave foresets (14-38°, ave. 20°), erosional lbs, commonly truncated ubs, or horizontal topset	L: v. thin to medium / m B: thin to thick / m to 10's of m BS: 1-1.5 m / 10's of m	 n=181	Subaqueous curved-crested dunes
Sp	Planar cross-stratified sandstone F-M quartz arenite, straight foresets (11-30°, ave. 19°), erosional lbs, commonly truncated ubs, or horizontal topset	L: thin to medium / m B: thin to v thick / m to 10's of m BS: up to 6 m / 10s of m	 n=97	Subaqueous 2D dunes
Sl	Low angle cross-stratified sandstone M-C quartz arenite, low angled (<15°) curved, convex and/or concave up, or planar foresets (Sl (p))	L: v thin to medium / m B: thin to medium / m to 10's of m BS: 30-50 cm (up to 1 m) / 10's of m	 Sl n=110  Sl (p) n=83	Transitional to upper flow regime subaqueous dunes or lat. accretion surfaces
Sr (a)	Unidirectional ripple cross-stratified sandstone F-M quartz wacke (rarely arenite), unidirectional small scale straight or curved foresets, sometimes mud draped, subcritical to supercritical climbing angle	L: v thin / cm LS: 0.5-3 cm / cm to m B: v. thin to thick	 n=45	Straight or sinuous-crested current ripples Unidirectional lower flow regime flows
Sr (b)	Oscillation ripple cross-stratified sandstone F quartz wacke with bidirectional or multidirectional small scale cross lamination, t common mud drapes, ripple index 5-15	L: v. thin / cm LS: 1-3 cm / cm to 10's of cm	 n=24	Combined current and oscillatory flows

Su	Low angled, undulatory cross-stratified sandstone Low angled wavy laminae, 20-50 cm wavelength interlaminated F quartz wacke and clay/silt	L = v. thin LS = 3–8 cm B = ≤ 20 cm	 n=10	Shallow water storm waves and currents
Sc	Sandstone rhythmite with mud drapes Subset of St, Sp, Sr and Sh containing mud draped laminae and mud couplets	Variable laminae and bed thickness Rhythmic lamination	 n=32	Incomplete tidal bundles
Sh	Horizontally-laminated sandstone Horizontally laminated F-M quartz arenite or quartz wacke sometimes with parting lineations	L: v. thin to thin / m to 10's of m B: medium / 10's of m	 n=2	Upper flow regime shallow flows
Sm	Massive sandstone Structureless beds (cryptic lamination) of M quartz arenite	B: Medium to v. thick BS: up to 3 m	no data	Debris flows and/or rapid deposition, water escape
SSDS	Soft-sediment deformation structures Occurs in any facies, parabolic c-s, chevrons, convolution, wrinkle marks, clastic dykes, faults, slumps	Variable	 n=14	Liquefaction by water escape, shock, or rapid changes in flow regime
Fh	Horizontally laminated shale and mudstone Gradational or sharp based, sharp ups, monolithic to heterolithic, or with thin sandstone interbeds, plant and trace fossils common	L = v thin to thick / 10's of m B = thin to v. thick / 10's of m to km	 n=3	Suspension settling and/or dynamic mud and/or flocculation

5.2. Sedimentary facies

5.2.1. *Intraformational conglomerate (Gi)*

Description– Thin to thick-bedded clast or matrix-supported intraclast conglomerate not exceeding a metre thick in the sections studied. Beds are erosively based, with deeply scoured (rarely straight) lower bounding surfaces and are generally limited in lateral extent, being developed locally in erosional depressions (Fig. 5.1A). Clasts are angular to round, often internally laminated and deformed at the margins. They are composed of every rock type found in the study area (laminated black shale, mudstone, siltstone and sandstone (quartz arenite and quartz wacke)), but most commonly of finer-grained rocks. Clasts are very poorly sorted, ranging from granule to medium boulder (boulder-sized clasts are common, but not ubiquitous to the facies). In rare cases clasts occupy a single size class, including boulders (Fig. 5.1B). Pebble to cobble-sized clasts are the most frequently recorded. Clasts are usually horizontal to sub-horizontal with respect to bedding, but may occupy a range of orientations including bedding perpendicular (Fig. 5.1C). The matrix is a range of grades of structureless quartz wacke. Certain examples blur the lines between clast and matrix, as the clasts can be interwoven and deformed in complex ways making them difficult to distinguish from matrix (Fig. 5.1D and E; Fig. 5.2A; Fig. 5.26E). Rounded sandclasts have, in various instances, been observed with an eye-shaped cross section with lateral pinch-out margins (Fig. 5.2A and C). One of the sand pebbles found *in situ* in the upper Rooirand Member was broken open, and contained a thin veneer of fine sand over carbonaceous black mud, which was therefore determined to be an armoured mudclast. Several other sandclasts appear to be composed of pure sand, as with the illustrated examples. Fossil plant axes are nearly ubiquitous to the facies, preserved as compressions or endocortical casts of isolated axes, rarely retaining branches or leaves (Fig. 5.2E and F). Most common among them are protolepidodendrolean lycopods, which have persistent leaves that are sometimes preserved even where the axis has

been filled internally by sediment. Arborescent lycopods bearing roots have been reported from this facies in sandstones underlying the Waterloo Farm Main Fish Lens (Prestianni and Gess 2014) and their axes were found in sections at Waterloo Farm and Seven Fountains in the upper Rooirand Member (Figs 5.23D and 5.28F). Fossil chondrichthyan spines occur in this facies at two localities in the ‘Skitterykloof Member equivalent’ near the top of the Witpoort Formation (Fig. 5.2G), but plant axes were not found in this member. The facies occurs at regular intervals throughout the Rooirand Member and in the Skitterykloof Member, but is not recorded in the Perdepoort Member. It is interbedded with structureless sandstone (Facies Sm). Facies Gi has been traced laterally to bedding-parallel stringers of intraclasts, the latter being far more common in the studied sections (Fig. 5.2B). Stringers are usually round to sub-round mud pebbles to cobbles, one to a few clasts thick, and are better sorted in comparison to the intraclast conglomerate. Clasts are ovoid to platy, usually elongated parallel to bedding, but sometimes show imbrication (Fig. 5.23C). Although attempts were made to measure primary current indicators from imbricated clasts and trends of clast elongation, the results were found to be unreliable and often contradictory within beds, and are not reported here. Intraclast stringers may be localised, limited to a few metres in extent or more extensive, being traced in places for tens of metres to the limits of outcrop visibility. They are hosted in almost every sandstone facies described below, typically lining the lower bounding surfaces, or, less commonly, along foresets of cross-stratified beds and are a regularly occurring phenomenon throughout the study area, except, as with facies Gi, in the Perdepoort Member.

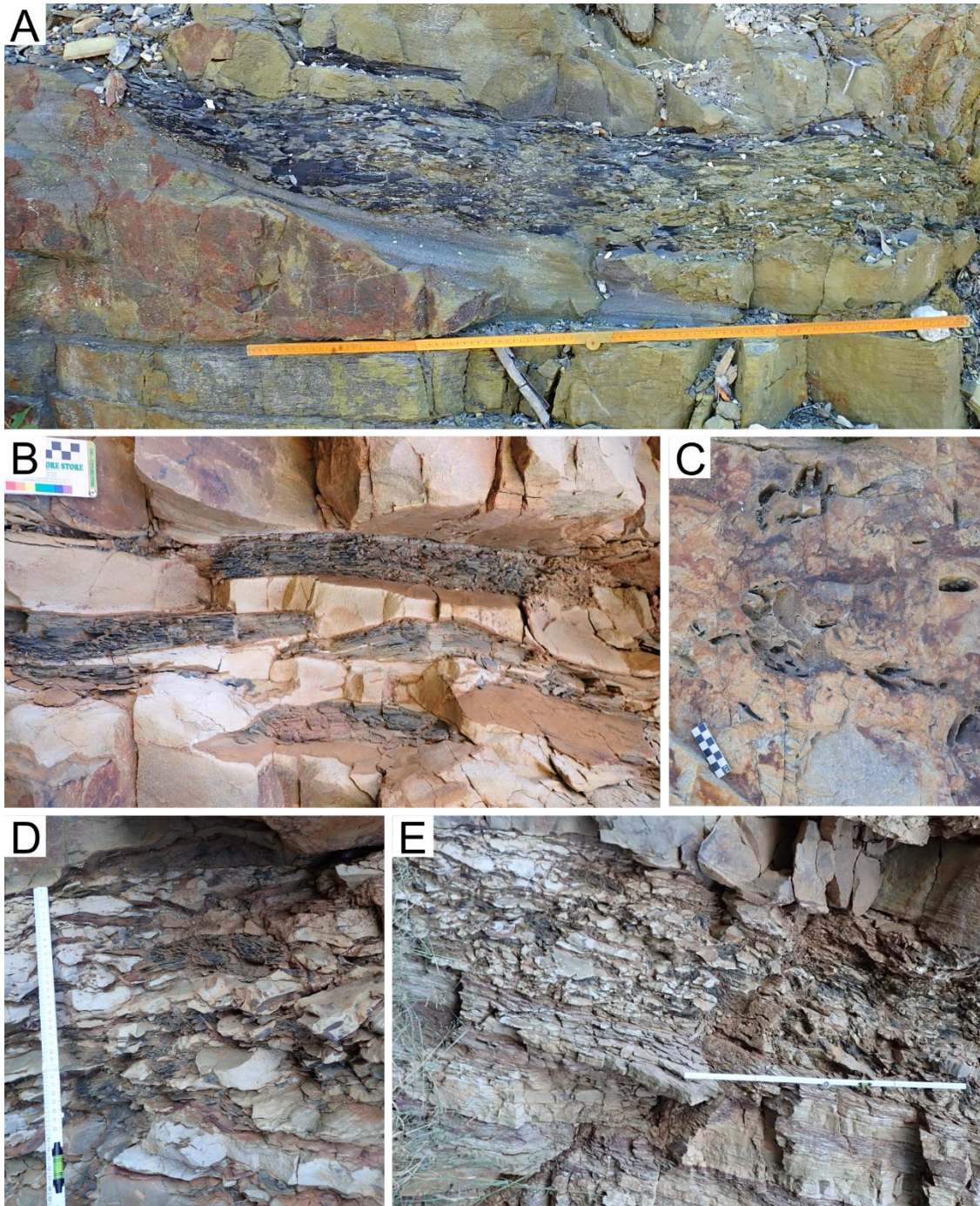


Figure 5.1: Intraformational conglomerate in the Witpoort Formation. A) Poorly-sorted shale clast conglomerate (granule to cobble) that is locally developed within a trough-shaped scour at Waterloo Farm (WF 1b). B) Boulder sized, deformed shale clasts embedded in a quartz wacke matrix at Waaipoort (WP 1). Clasts at this horizon are up to 64 cm long. C) Platy mudclasts that are emplaced at various angles including vertical (top of photograph) from Howison's Poort (HP 2). D and E) Thick bed with complexly interwoven, deformed sandclasts and mudclasts. D is an enlargement from the left of the photograph in E. Rulers in A and E represent 1 m. Scale increments in B, C and D are 1 cm.

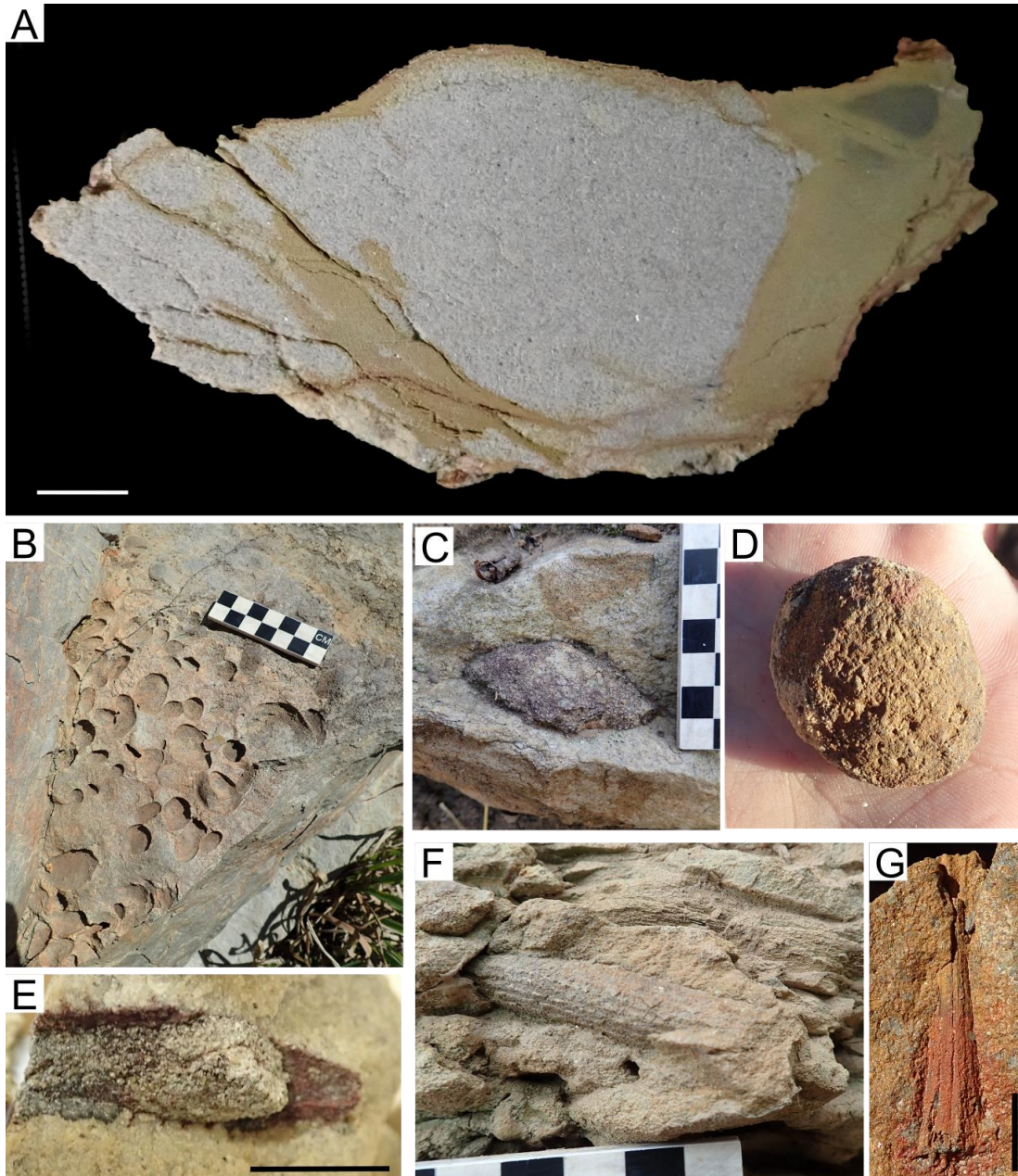


Figure 5.2: Intraformational conglomerate clasts and bioclasts in the Witpoort Formation. A) Cross section of a clast-supported conglomerate from Howison's Poort (HP 2) with a deformed cylindrical sand clast with lateral pinch-out margins cut transversely. The clast is surrounded by a veneer of claystone and other deformed sandclasts (scale bar = 1 cm). **B)** Bedding-parallel stringer of rounded, weathered out mud pebbles lining the lower bounding surface of a cross stratified bed at Waterloo Farm (WF 2). **C)** Sand clast from HP 4 in profile view showing its eye-shaped morphology with lateral pinch-out at the margins. The clast is coarser grained than the sandstone matrix. **D)** Sand pebble from Seven Fountains, about 4 cm in diameter. **E)** Endocortical cast of a protilepidodendrolean lycopod with spine-like leaves attached. **F)** Lycopod axis with leaf scars from the bed illustrated in A. **G)** A putative chondrichthyan fin spine from the uppermost Witpoort Formation at Alicedale. Scale bars in A, E and G are 1 cm. Scale increments in B, C and F are 1 cm.

Interpretation– The clasts include a variety of sedimentary rock types that are recorded in the overlying and underlying deposits, and it is likely that they originate proximally within the depositional basin. The existence of lamination in many of the clasts, particularly the shale clasts, indicates some prior consolidation, but the deformations recorded here show that the clasts remained soft during transport and burial. This facies may arise in any depositional environment where downcutting with rapid transport and dumping of clasts of semiconsolidated sediment occurs (Dietrich *et al.* 2017). Similar facies in the rock record may represent regular sedimentary processes such as channel lags (Cant and Walker 1976; Miall 1977; Godin 1991; Bridge 1993) and slumps (Plint 1986; Aschoff *et al.* 2018), or may be associated with changes in base level, marking regionally significant stratigraphic surfaces (Miall and Arush 2001) and separating these mechanisms can be challenging. Intraclasts associated with massive sandstone and soft-sediment deformation in ancient strata have been attributed to slumping (Jones and Rust 1984), and this possibility is supported by variably orientated clasts, including originally vertical, at certain horizons, as well as the strong association with structureless sandstone (Facies Sm).

Lateral grading from conglomeratic beds to stringers of intraclasts indicates a similar process origin for the latter, which, however, have better rounded and sorted clasts indicating greater degrees of transport. Parallel-laminated shale clasts suggest erosion from previously deposited mud beds. Platy clasts would likely have been sourced from laminated, semi-consolidated shale, whereas ovoid clasts indicate a weakly laminated or structureless mudstone source (Bridge and Demicco 2008). Some idea of their potential distance of transport is derived experimentally. Flume tank studies show that semi-consolidated mud can be eroded by current stronger than 16 cm/s (Schieber *et al.* 2010). Mudclasts transported in a racetrack flume tank were rapidly rounded during transport and destroyed within tens to hundreds of metres (Smith 1972). Smith, however, acknowledged that his experiments did

not account for factors like armouring or algal binding of clasts (Smith 1972). More recent flume experiments (Schieber *et al.* 2010), however, have shown the potential distance of mudclast transport can exceed the estimates of Smith (1972) by orders of magnitude. Though both of these studies were performed in racetrack flumes, a key discrepancy between the studies is the construction of a sandpaper bottom in the earlier study, and may better approximate conditions of mudclast transport in the Witpoort Formation. Surface roughness is considered an important variable increasing the abrasion and thus reducing the potential transport distance of mudclasts. Additional factors influencing the durability of mudclasts include prior consolidation, the velocity of the transporting current and armouring (Schieber *et al.* 2010; Hizzet *et al.* 2020). In the absence of armouring, many of the clasts may have originated within the immediate depositional environment with transport distances on the order of tens to hundreds of metres, as arrived at by Smith (1972).

Transport and rounding of sandclasts requires a mechanism of sediment binding, since compaction of sand alone is unlikely to produce cohesiveness sufficient for transport (Pflüger and Gresse 1996; Bridge and Demicco 2008). Post-depositional deformation of the clasts is apparent, in particular the pinching at the margins, indicating that the clasts were malleable at the time of burial. Two explanations for their pre-transport consolidation are envisaged: algal binding or freezing of soft/semi-consolidated sediment (Pflüger and Gresse 1996). A third explanation, that of early cementation of the clasts is unlikely because the clasts would then be expected to have remained hard after burial (Runkel *et al.* 2010). Sand intraclasts are common in periglacial deposits (Diffendal 1984; Menzies 1990), related to subglacial deformation (see Waller *et al.* 2011), or glacial outwash, in which clasts may be transported in a frozen state (Reinson and Rosen 1982; Runkel *et al.* 2010; Dietrich *et al.* 2017). Algal binding has been suggested to have enabled the transport of cohesive unlithified sand clasts in Precambrian strata: a time when algal mats were prolific (Pflüger and Gresse 1996). Evidence

of algal binding in the study area is limited to uncommon wrinkle structures on bedding surfaces, where algal mats appear to have been less predominant. The hypothesis that sandclasts were transported frozen is preferred, considering other evidence for periglacial processes in the Witpoort Formation (Almond *et al.* 2002), such as extraformational conglomerate (5.2.2), dropstones (5.2.10) and certain soft-sediment deformation structures (5.2.12), which potentially reflect periglacial processes, in light of the stratigraphic evidence and palaeogeographic setting (see Chapter 8.6).

5.2.2. *Extraformational conglomerate (Ge)*

Description– The facies comprises matrix-supported lithic clasts emplaced in cross-stratified, medium to very coarse-grained quartz arenite and quartz wacke. The clasts occur as stringers lining the upper and lower bounding surfaces of beds and along laminae (Fig. 5.3A). They range from granule to coarse pebble in the study area and can occupy a range of size classes or a single size class at any given horizon (Fig. 5.3 B and D). Clasts are subangular to round, predominantly opaque and translucent white quartz, but also include fine-grained shaly clasts and probable sand clasts.

In bedding plane expression the clasts occur in localised concentrations on the upper bounding surfaces of low-angle cross-stratified (Facies SI) beds. They may occupy bedding-parallel zones of decimetres to metres in extent with the remainder of the bed surface lacking clasts. At one horizon (WP 2) a linear concentration zone of clasts was found to be elongated along the same trend as the dip of foresets in the underlying bed.

The facies is limited to the Skitterykloof Member at the top of the Witpoort Formation, and is only found at Waaipoort in the studied stratigraphic sections. Multiple conglomeratic horizons occur sporadically in the member. Westward of the study area, the facies is increasingly prevalent, forming thicker accumulations and including larger clast sizes, up to coarse cobble (Fig 5.3C).

Interpretation– Though conventional facies analyses do not always discriminate between conglomerates with clasts that were lithified or not at the time of deposition, this characteristic appears to be significant in the case of the Witpoort Formation. Lithic pebble conglomerate is a commonly recognised facies in ancient successions, often occurring in lag deposits and gravel bars (e.g. Allen 1983; Miall 1985; Reesnik and Bridge 2007; Jinnah and Roberts 2011). Their association with low angle cross-stratification in the study area indicates high-energy depositional conditions, consistent with conventional explanations for this facies. Concentrations of clasts lining the upper bounding surfaces of sandstone beds potentially represent exposed surfaces where deflation by wind has concentrated larger particles (Collinson and Mountney 2019). Occurrence of lithic clasts at only one stratigraphic interval near the top of the Witpoort Formation indicates a significant forestepping of the depositional locus. This is the first appearance of lithic conglomerate in the Witteberg Group. In underlying strata, lithic clasts are recorded in the Pakhuis Formation (Table Mountain Group), deposited during the Hirnantian (end-Ordovician) glaciation (Blignault and Theron 2010), potentially indicating a comparable explanation. Clasts were examined for diagnostic features of glacial transport such as facets and striae (Bridge and Demicco 2008), but no clear evidence of this was found.

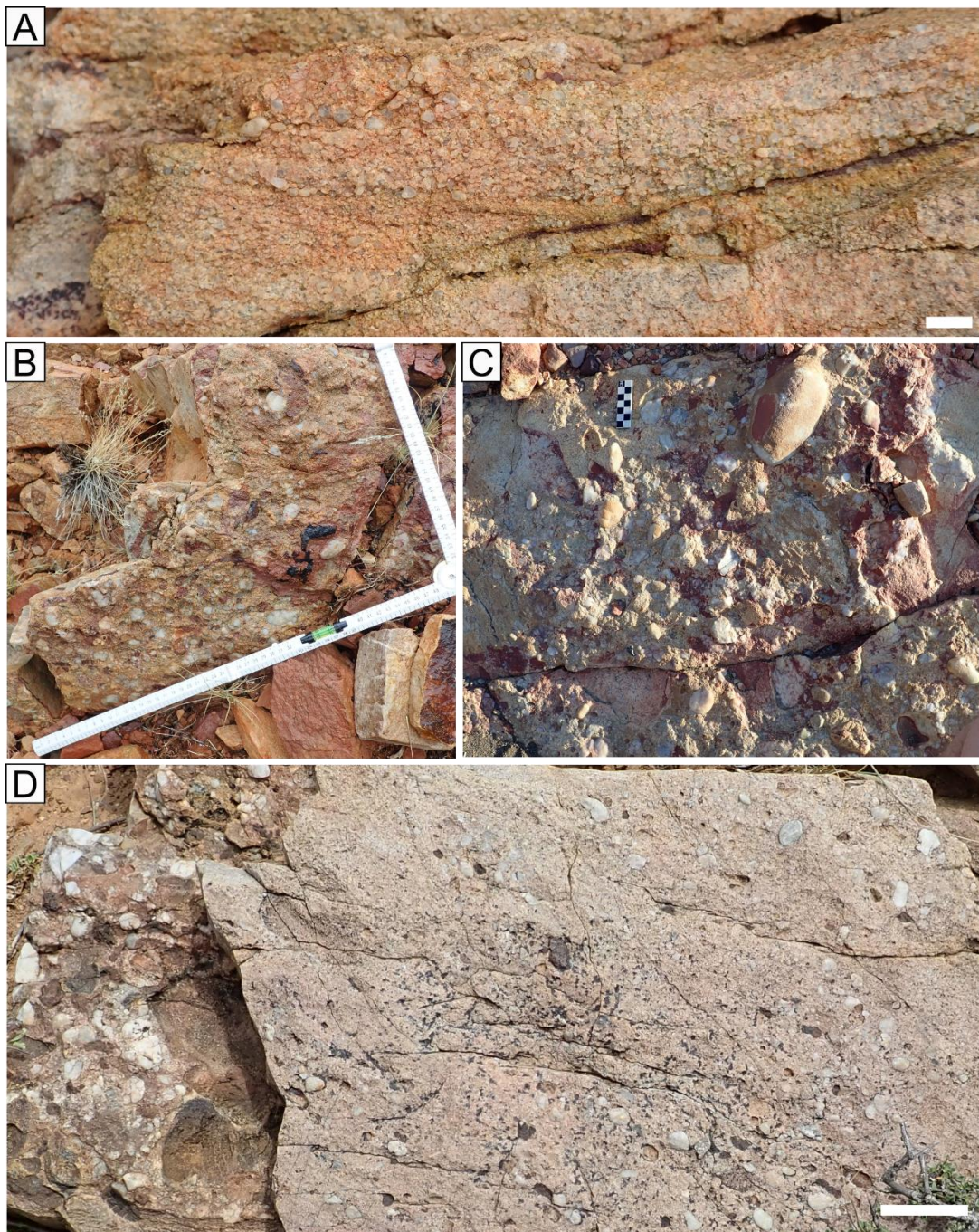


Figure 5.3: Extraformational conglomerate in the Witpoort Formation (Skitterykloof Member). A) Profile showing stringers of quartz granules emplaced along low angle cross-laminae contacts (scale bar = 1 cm). B) Poorly-sorted conglomerate lining the lower bounding surface of a bed. C) Upper bounding surface of a sandstone bed at Skitterykloof (Western Cape) with clasts up to cobble-sized. D) Variable size and sorting of clasts is evident in this bed (scale bar = 5 cm).

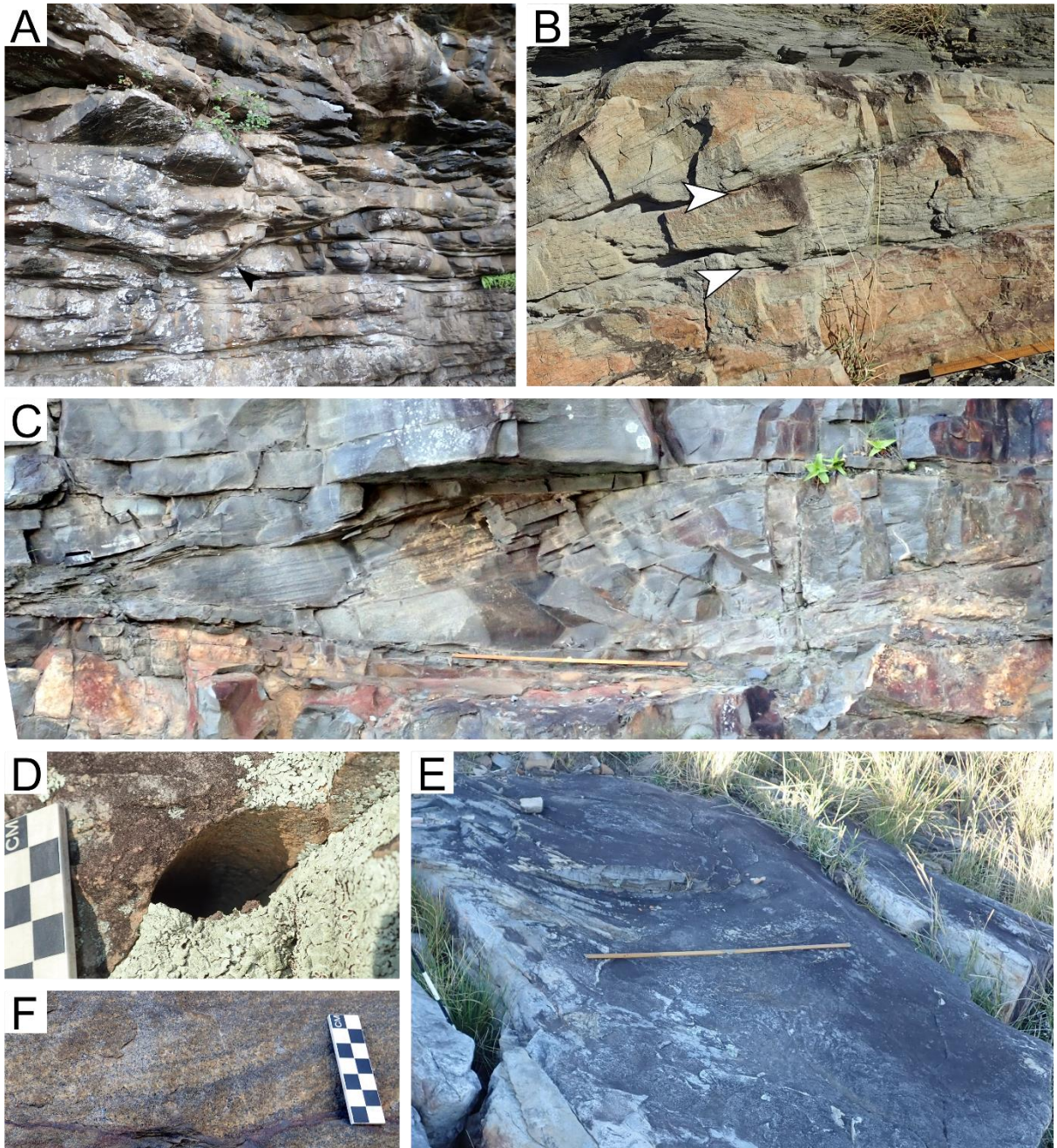


Figure 5.4: Trough cross-stratification (St) in the Witpoort Formation: A) An erosive channel cutting with more than a metre of downcutting relief exposed perpendicular to flow direction from HP 3. Erosion surface is arrowed. B) Lenticular St with reactivation surfaces (arrows) from WF 2. Note that St is overlain by a thin Sh topset, followed by horizontally laminated shale (Fh) at Howison's Poort (HP 3). (Ruler segment is about 15 cm long). C) A thick, sigmoidal St bedset that shallowly underlies Fh (Ruler is 1 m long). D) Hollowed-out plant axis in St at the top of WF 2 (scale in cm). E) Bedding plane exposure with large scale rib and furrow structures on the top surface of the bed containing the fossil in D (Ruler is 1 m long). F) Small scale St with subtle heavy mineral laminae that define foresets that are strongly tangential to the lower bounding surface of the bed (scale in cm).

5.2.3. Trough cross stratified sandstone (St)

Description– Trough cross-stratified sandstone is characterised by curved lower set bounding surfaces (LBSs) and inclined, upwards-convex foresets. This facies comprises foresets that are inclined at a moderate angle (15–35°, ave. 19°, n=215) and are strongly tangential or concordant to the LBS (Fig. 5.4). Flow vectors for this facies are unimodal to SW (Table 5-1). Facies St is developed in light blue-grey, olive grey and steel grey, medium- to coarse-grained, mature quartz arenite in the study area. Two end members of this facies exist, they are very laterally restricted, lenticular forms and laterally extensive, tabular forms. The first group tends to form thin to very thick (up to 110 cm thick), lens-shaped or sigmoidal beds that pinch out at a scale of metres to tens of metres. Laminae are defined by heavy mineral laminae (Fig. 5.4F), mud drapes and/or sharp textural contrasts between laminae, producing very well-defined laminae partings (Fig. 5.4B and C). Heavy mineral laminae do not form parting planes in the rock, whereas clay drapes tend to be negatively weathered and cause weaknesses that cause the rock to part. Reactivation surfaces are common (Fig. 5.4B), particularly in the more lenticular forms of St. Reactivation surfaces are referred to in the sense outlined by Martinius *et al.* (2015), which include slight changes to foreset orientation corresponding to minor erosion surfaces within a bed. The lenticular form of St occurs in lithologically variable successions with interbedded ripple cross laminae and ripple form sets, mudstone and shale, rib and furrow structures (including centimetre and metre scale forms; Fig. 5.4E) and, in places, plant and trace fossils (Fig. 5.4D). Cross strata toe sets have accumulations of fines and ripple cross-lamination. By contrast, the second form is more monolithic and amalgamated, and is generally more texturally mature. It forms medium to thick trough cross-stratified beds that are amalgamated as bedsets, usually on the order of 1–1.5 m thick, which extend for tens of metres along strike. This latter form tends to have a supermature texture and a more massive appearance with more or less distinct heavy mineral

laminae developed along foresets, which have a weak or non-existent parting (Fig. 5.4F). It forms lithologically monotonous successions, associated with structureless and low-angle cross-stratified beds. Ferruginous, or slightly muddy partings along bedset contacts, commonly with back-climbing ripple cross laminae, form slight notches that separate the tabular strata. Plant and trace fossils are rarely associated with the amalgamated, tabular form of St. Intraclast stringers as well as isolated ‘floating’ mudclasts are regularly associated with both forms of St. Patterned surface textures are common on planar and trough foresets in the upper Rooirand Member and lower Perdepoort Member in the Seven Fountains section. The textures include low-relief ridges arranged in irregular linear or interference patterns and are sometimes superimposed with small wart-like structures (Fig. 5.5 A–E). The features are interpreted as adhesion ripples and warts.

Interpretation– Trough cross-stratification forms by the migration of sinuous crested, three-dimensional dunes (Miall 2006). Dunes are defined as bedforms greater than 5 cm in height (Ashley 1990). Curved crested dunes form under higher current velocities than straight-crested dunes and are indicative of relatively high-energy, lower flow regime conditions (Froude number < 1) (Reineck and Singh 1980; Miall 2006; Nichols 2009). Avalanching of grains on the lee (downstream) face is the dominant mechanism of dune progradation (Allen 1982). Avalanching occurs where the lee slope reaches a critical angle due to accumulation of grains on the dune crest, known as the angle of repose (30–40° according to Bridge and Demicco 2008). The average St foreset inclination of 19° is well below this range. Backflow due to turbulence in the dune troughs leads to the tangentiality at the base of foresets and can produce backflow ripples which prograde up foreset dip and co-flow ripples that prograde in the current direction (Bridge and Demicco 2008; Collinson and Mountney 2019) as well as the accumulation of fine material along toe sets (Reineck and Singh 1980). Backflow and grainfall were more important processes in the formation of the

lenticular St described above, whereas the amalgamated form lacks significant accumulations of fine-grained material and rippling, which may not preserve under certain conditions (Martinius *et al.* 2015). Heavy mineral laminae, which are the primary form of lamination in the amalgamated form, arise due to their greater density and resistance to resuspension relative to lighter minerals (Cheel and Middleton 1986). Important controls on the scale of subaqueous dune forms are water depth, sediment grain size and flow velocity (Allen 1982). Subaqueous dunes that have heights in metres tend to form in deeper flows, greater than 5 m and up to tens of metres deep (Reineck and Singh 1980; Leeder 1983; Mazumder 2003; Nichols 2009). In general, dune height increases with water depth (Allen 1982; Mazumder 2003; Best and Fielding 2019). It is likely, therefore, that thick to very thick-bedded forms would have formed under deeper flows than thin to medium beds, however, the inherent problems with attempting to quantify flow depth from preserved dune height are discussed by Martinus *et al.* (2015). Dunes can form in subaerial and subaqueous settings such as rivers, estuaries and tidal flats, shorefaces and open shelf settings (Allen 1982; Nichols 2009). Dunes deposited by wind tend to show inverse grading of foresets, steeply inclined, thick foresets, irregular cross-set thicknesses and greater scales of cross-stratification than subaqueous forms (Evans 1944; Reineck and Singh 1980; MacNaughton *et al.* 2002; Nichols 2009), as well as an abundance of slump structures (Reineck and Singh 1980). The moderate angles of foreset dips and relatively uniform medium to thick-bedded nature of this facies is consistent with subaqueous rather than subaerial dunes. However, possible exposure of the beds is insinuated by the presence of what have been interpreted as adhesion marks, which form due to wind-blown sand adhering to a wet surface, according to wind tunnel experiments (Reineck 1956; Kocurek and Fielder 1982). They are also referred to as antiripplets because the inclination of the ripple foresets is up-current (Reineck and Singh 1980). Adhesion warts are formed due to rapidly changing wind direction, which inhibits the

ripples from developing (Reineck and Singh 1980). In the study area they have only been found in planar and trough cross-stratified facies. This bedding surface texture was recognised in the Seven Fountains section late in the study, but whether it occurs at other horizons and sections is uncertain, as similar rugose textures had been interpreted as a structural artefact in other sections where bedding surface exposure is often far more limited. Flow vectors for St (n=181, Table 5-1) have a strong SW modal orientation consistent with progradation and downstream accretion, although superficially similar forms of trough cross-stratification can form in a variety of subaqueous depositional environments and their discrimination relies upon facies stacking patterns, lateral relationships and fossil content, which are further explored below (5.4).

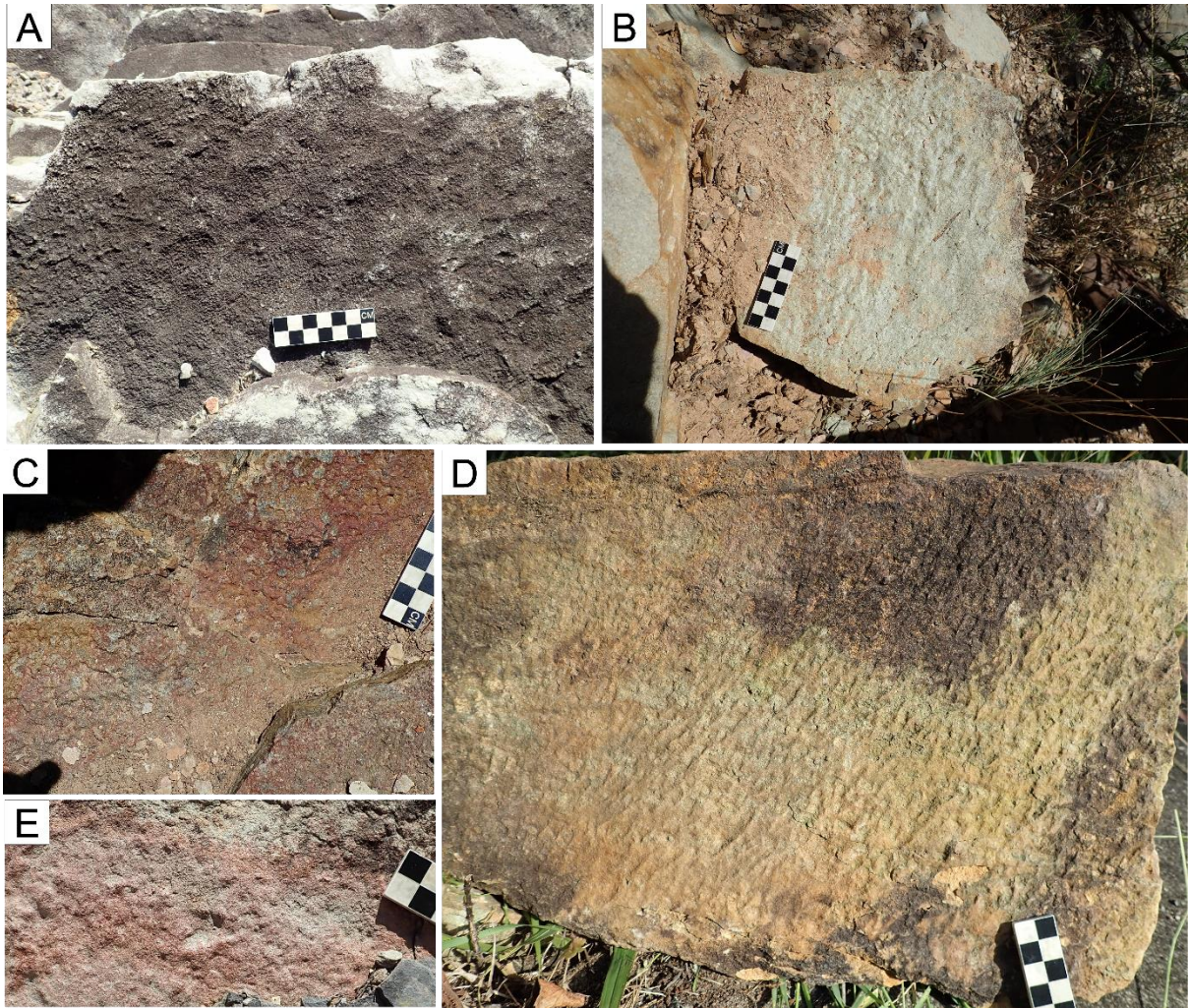


Figure 5.5: Adhesion ripples and warts (All scale increments are 1 cm). A) Low relief adhesion ripples with superimposed warts, flow is to bottom right of photograph. B–D) Adhesion ripples on slightly mud-draped foresets. Flow in D is to left of photograph). E) Possible adhesion warts on a mud-draped flat bed.

5.2.4. Planar cross-stratified sandstone (*Sp*)

Description– Planar cross-stratification is defined by planar, inclined foresets deposited upon a flat lower bounding surface. It is usually developed in fine to medium-grained, mature, light grey, olive grey and light blue-grey quartz arenite. Foresets are thin to medium laminated and are inclined on average at an angle of 23° relative to the lower bounding surface ($n=97$). The orientation of *Sp* foresets in the study area is very similar to that of *St* with a unimodal trend to SW. Foresets are separated by heavy mineral laminae, thin clay linings or cryptic textural discontinuities between adjacent laminae (Fig. 5.6B and D). *Sp*

beds are wedge-shaped or tabular bedforms ranging from thin to very thick bedded. Foresets are normally truncated at the upper bounding surface of a bedset and are underlain by planar surfaces of erosion. Reactivation surfaces are common in this facies, particularly in medium to thick bedded, laterally restricted examples. Commonly a drape of fine-grained material is developed along bed or bedset contacts. Beds are 5–40 cm in thickness, although rarely may attain much greater thicknesses (the thickest bed recorded in the study area is 1.6 m thick) (Fig. 5.6A). Multiple similar beds may be amalgamated within bedsets from 30 cm up to several metres thick (maximum 5–6 m), within which beds thin and fine upwards (Fig. 5.6B). Cross stratification is generally unidirectional within a bedset (Fig. 5.6A, B and E). Sp is less abundant than St in the Witpoort Formation.

Interpretation– Planar cross-stratification forms under similar hydrodynamic conditions as trough cross-stratification (5.2.3 above), but differs in representing straight-crested rather than sinuous crested dunes. Straight-crested, two-dimensional dunes, under relatively lower current speeds than the latter, all other variables being equal (Reineck and Singh 1980; Ashley 1990; Collinson *et al.* 2006). According to these studies, greater water depth and/or larger grain sizes could favour planar cross-stratification, explaining the thicker beds that characterise this facies compared with trough cross strata. These variables aside, the conditions of genesis may be similar for both facies in the Witpoort Formation, since the geometry, foreset angles and lithology are comparable. Thus, according to Ashley (1990), large-scale flow-transverse bedforms that produce planar and trough cross-stratification form a continuum rather than discrete categories.



Figure 5.6: Planar cross-stratified quartz arenite (facies Sp) (ruler is 1 m long). A) An upward thinning Sp bedset overlain by a 1.6 m thick tabular Sp bed. Stratification is outlined in the photograph. B) A 1 m thick, amalgamated Sp bedset with a thick tabular bed overlain by medium to thin beds. Stratification is unidirectional within the bedset. Small arrows indicate bed contacts, larger arrows indicate the bedset contact (segment of ruler is 50 cm long). C) Thick Sp bed with well defined, heavy mineral laminae, which are enlarged in (D), where the heavy mineral accumulations can be seen towards the tops of each medium lamina. E) A 1.5 m thick Sp bedset formed by wedge-shaped thick beds passing upwards to tabular medium beds. Slight silty accumulations are seen along bed lower bounding surfaces. Laminasets of cross-strata, separated by slight reactivation surfaces, show a strong parting due to the textural discontinuities at those contacts. Stratification is unidirectional within the bedset.

5.2.5. *Structureless sandstone (Sm)*

Description– Structureless, medium to thick beds of texturally mature quartz arenite are pervasive in the study area (Fig. 5.7A–C). They form tabular beds or extensive lenses, normally extending for at least tens of metres laterally. Lamination is not detected in this facies. In places, however, apparently massive strata were traced laterally to St with faintly detectable heavy mineral laminae. In some cases laminae can be detected in the lower parts of otherwise apparently massive beds due to slight clay accumulations along the toe sets of these beds. This implies that some of the massive strata contain cryptic lamination, an effect which may be exacerbated by weathering, structural deformation or mineral precipitation (Fig. 5.7A). Facies Sm is associated with facies Gi in many examples in the study area, normally occupying the strata overlying these intraclast horizons. It is also interbedded with cross-stratified facies, to which it is geometrically similar, where a Sm bed is probably the geometric equivalent of an St or Sl bedset. Thin to medium, tabular structureless beds are also recorded within heterolithic successions. Subtle soft-sediment deformation structures are detected in some apparently massive beds (Fig. 5.7C).



Figure 5.7: Thick bedded structureless quartz arenite facies (Sm). A) The bed is influenced by jointing and mineral precipitation in the fractures, no lamination is visible. B) Sm is seen at the top of the photograph, overlying a thin conglomeratic horizon (Facies Gi, arrowed). Below this horizon a sandstone bed with parabolic soft-sediment deformation is seen at bottom of photograph (scale bar represents 50 cm). C) The upper bounding surface of this bed is affected by soft-sediment deformation, which may have also destroyed prior lamination in this apparently massive bed.

Interpretation— Massive sandstones can arise in three ways; either from the absence of primary lamination; from its subsequent destruction; or by having laminae that are invisible at outcrop (cryptic lamination). For an example of the latter, Cheel and Middleton (1986) recorded a variety of modern and ancient examples where, in the absence of heavy mineral laminae, strata appeared structureless in outcrop and thin section, but deduced that the strata contained cryptic lamination. This effect can be exacerbated by weathering and

mineralisation in fractures in the rock that obscure laminae (Fig. 5.7A). Truly structureless beds have been attributed to rapid sedimentation by mass flows (Miall 1977, 2010) and hyper-concentrated flows (Russel and Arnott 2003; Davies and Gibling 2010). Although these terms may be applied interchangeably, mass flows and hyper-concentrated flows exist on a continuum from bedload to suspension-dominated transport (Calhoun and Clague 2017). They are related to upper flow regime unidirectional flow, likely during peak flood conditions (Russel and Arnott 2003). This may explain some of the massive strata in the study area, particularly those associated with intraclast conglomerate and affected by fluidisation (5.2.12). It is not excluded, however, that some of the structureless strata contain cryptic lamination, an effect which is exacerbated by structural deformation and weathering in the study area. Destruction of laminae by cryptic bioturbation is, however, ruled out as no trace fossils were recognised along bedding contacts or in polished profile sections of this facies (Greene *et al.* 2012). Destruction of primary lamination by fluidisation is a likelihood (Jones and Rust 1983; Collinson and Mountney 2019), since some of the massive strata are associated with soft-sediment deformation structures. It is therefore plausible that a combination of cryptically-laminated and genuinely structureless sandstone are included in this facies.

5.2.6. *Low angle cross-stratified sandstone (Sl)*

Description– Low angle cross-stratification occurs in two forms, low angle trough cross-stratification (Sl), and low angle planar cross-stratification (Sl (p)). They are separated by having curved or planar foresets, which are inclined at an angle of less than 15° relative to bedding. Sl and Sl (p) are common throughout the Witpoort Formation. This facies is texturally mature, medium- to coarse-grained quartz arenite or sub-arenite. It is developed in medium (rarely thick) beds, which are amalgamated in tabular bedsets. In this regard the facies has a very similar outcrop expression to tabular St described above and the two facies

are often mistaken. They are separated based on the angle of foreset dip, which is placed at below 15° for Sl (Miall 2006). Upon back-rotating the foreset orientations, it was found that 44% of beds coded in the field as Sl contained foresets dipping at more than 15°, and conversely, 28% of St beds resolved at below 15°. Wherever a rock profile exposure is not parallel to the foreset dip direction the apparent dip will be less, leading to an overrepresentation of Sl in the field logs (Appendix 1C–L). Nonetheless, these two facies appear very similar in the field, and can be interbedded within the same bedset. The main distinction between Sl and St, apart from the foreset angle, is the polymodal foreset orientations of the former, versus the unimodal of the latter (Table 5-1). Sl foresets have multidirectional dips, they are concave or convex up, concordant or discordant to the lower bounding surface, parallel-laminated or slightly divergent. Sl (p) is very similar in outcrop expression, except that foresets are more unidirectional within a bedset, and present a unimodal distribution (Table 5-1), and the bed or bedset bounding surfaces are flatter.

Laminae are very thin to thick, marked by very thin, sharp or indistinct heavy mineral linings, seldom with clear textural discontinuities, normally with a weak or non-existent parting (Fig 5.8). Laminae are arranged in lenticular medium beds (usually less than 30 cm thick) separated by erosive contacts. Beds are amalgamated in laterally extensive bedsets with thicknesses of 0.8–2.5 m, that extend for up to several tens of metres along strike. They have several metre long undulations, scours and hummocks along the lower bounding surface (Fig. 5.8). The bedset is the most visible form of stratification in this facies, their lower bounding surfaces commonly lined with very thin accumulations of fine-grained material, and sometimes with submature, fissile toe-sets containing back-climbing or downclimbing, irregular ripples. These linings of finer grained material form slight notches in the outcrop along bedset contacts. Facies Sl and Sl (p) are the predominant facies in the Perdepoort Member, but occur throughout the formation.



Figure 5.8: Low angle cross-stratified quartz arenite (Facies Sl). Above: Low angle planar cross-stratification (facies Sl (p)) in the Perdepoort Member at Waaipoort (WP 2) (jacob staff is 1.5 m). Below: Low angle cross-stratification with curved laminae (facies Sl) (metre rule for scale) from Seven Fountains. Observe amalgamated bedsets of approximately a metre thick in both photographs.

Interpretation– This facies closely resembles medium-scale St in the current normal view. They are texturally similar, and co-occur in thick arenitic successions. They are distinguished in that Sl laminae are low angled in three-dimensions, whereas St laminae are steeper in the current parallel view. In some instances, however, measurements of Sl foresets showed them

to be steeper than interpreted at outcrop. Low angle cross-stratification forms under high energy conditions, such as under storm waves in shoreface and lacustrine settings (e.g. hummocky and swaley cross-stratification) and under high energy unidirectional currents in foreshore, estuarine and alluvial settings. Differentiating storm wave from current-produced, low angle cross-stratification is subtle and relies upon the stratigraphic context of strata, the grain size and the foreset orientations (Harms *et al.* 1975; Dott and Bourgeois 1982; Rust and Gibling 1990; Alexander *et al.* 2001). The range of grain sizes for Sl exceed those expected in hummocky cross-stratification, which is furthermore typically associated with small oscillation ripples and marine trace fossils (Harms 1975). There is, however, no evidence supporting a marine origin for this facies in the Witpoort Formation, as discussed below. Examples of this facies in ancient and recent fluvial deposits are numerous (e.g. Allen 1983; Allen and Underhill 1989; Rust and Gibling 1990; Cotter and Graham 1991; Miall 2006; Fielding 2006). In these settings, low angle cross laminae form in washed-out dunes and humpback dunes under conditions transitional between subcritical and supercritical flow, and in antidunes, chutes and pools, indicative of upper flow regime current exceeding the upper stage plane bed field (Alexander *et al.* 2001; Miall 2006; Fielding 2006). Characteristic bedforms of antidunes, chutes and pools are low angled, upstream-dipping laminae and downstream-dipping steeper laminae (Alexander *et al.* 2001). An important criterion in the distinction of these bedforms is the orientations of cross-laminae and truncation surfaces relative to flow (Fielding 2006). Structural deformation of strata makes such evaluations at outcrop challenging in this study area. Palaeocurrent vectors from this facies indicate both upstream and downstream sediment transport, and it is likely that both transitional and upper flow regime bedforms are encompassed by this facies. Based on laboratory studies, Miall (2006) considered antidunes to have had a low preservation potential in the ancient record. Such upper flow regime bedforms have the highest preservation potential in the falling stages

of a discharge event, and are most likely to form in rivers with highly seasonal, flashy discharge (Fielding 2006).

Though different processes are inferred for the low angle cross-stratification, this facies actually forms a continuum with the dip angles of planar and trough cross strata in the study area, as indicated by a plot of the dip angles recorded in all of these facies (Fig. 5.9A and B). Their separation at the angle of 15° (cf. Miall 2006) is therefore arbitrary and does not correspond to natural populations in the Witpoort Formation, but represent a continuum of depositional conditions. This continuum of low to high-angled dune foresets mimics the ranges of lee slope angles from modern-day large rivers (Cisneros *et al.* 2020).

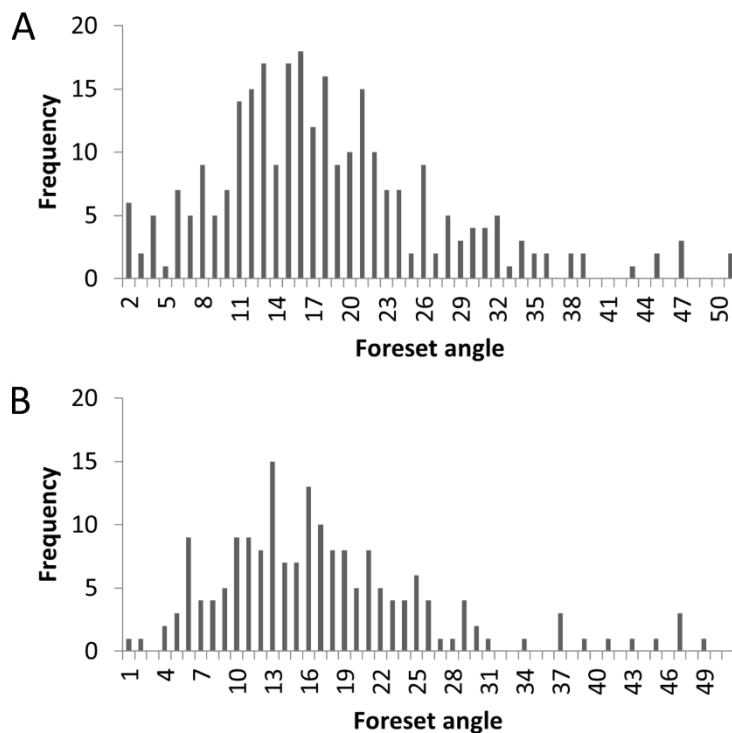


Figure 5.9: Cross-strata foreset dips from this study. A) Curved foresets (facies St and Sl). B) Straight foresets (facies Sp and Sl (p)). Both curved and straight foresets in the study area form continua and the 15° separation between facies is therefore arbitrary.

5.2.7. *Undulatory sandstone and mudstone (Su)*

Description– Su is characterised by low-angled, gently curved, upwards-convex and concave laminations that can undulate evenly (Fig. 5.10B). It is formed by parallel to sub-parallel, very thin to thin-laminated, light brown, grey or olive grey, fine-grained quartz wacke. Usually, the facies has a heterolithic aspect drapes and interbeds of fine-grained sediment that produce a streaky appearance. The laminae can be developed in rhythmic laminasets that are 2–8 cm (up to 15 cm) thick, separated by very slight erosional discordances and/or mud drapes (Fig. 5.10A and C). Discordances can often be traced laterally to a conformable surface. Wavelengths of undulations are recorded at between 20–50 cm with an amplitude of 1–3 cm. Longer wavelength forms occur but the continuity of undulations is usually interrupted where it exceeds the recorded wavelengths. Su typically grades vertically and laterally to flat-laminated sandstone (Sh) with which it is intimately associated. Interbedding with laminated mudrock and ripple cross-laminae is ubiquitous. Beds of this facies are thin to medium, having a pinch and swell aspect. They may be amalgamated in bedsets not exceeding a metre in thickness.

Interpretation– Facies Su resembles Sl somewhat, although it has a more undulatory aspect and comprises finer-grained, immature strata with ubiquitous muddy intercalations. This facies has some characteristics that point to oscillatory flows. Some typical features of hummocky cross-stratification characterise this facies (Dumas and Arnott 2006; Perillo *et al.* 2014), which include very low angle discordances that are laterally conformable, sub-horizontal wavy laminae and regular association with inferred combined flow ripple marks (5.2.9. Facies Sr (b)). Hummocky cross-stratification (HCS) is typically deposited by storm waves, although it is not in itself diagnostic of the palaeoenvironment; it can occur in the shoreface and surf zone, tidal flats, open-mouthed estuaries and lakes (Greenwood and Sherman 1986; Clifton 2006; Yang *et al.* 2006, 2007; Dumas and Arnott 2006; also see

Campbell and Oaks 1973). If Su does indeed represent HCS then it occupies the lower end of the range of wavelength expected for hummocks, which is 0.1 m to 10 m (Craft and Bridge 1987; Banerjee 1996). Wavelengths of oscillation ripples and hummocks form a continuum related to wave orbital diameter, which increases with shallowing in the shoreface and decreases landwards from the shoreline (Yang *et al.* 2006). Comparable wavelengths to those measured here (30–50 cm) have been reported from open-coast tidal flats and within a tidal estuary in South Korea (Yang *et al.* 2006, 2007), in the latter case about 1 km landwards of the mouth. To properly understand the conditions of deposition for Su, concerted attention should be given to its variability in wavelength, petrography, and vertical and lateral facies changes. The intimate relationship with muddy strata and gradation to horizontally laminated sand imply that, in addition to possible storm waves, high energy flows (5.2.8) and elevated suspended sediment concentrations were important controls on Su (see 5.4.2 for further extrapolations).

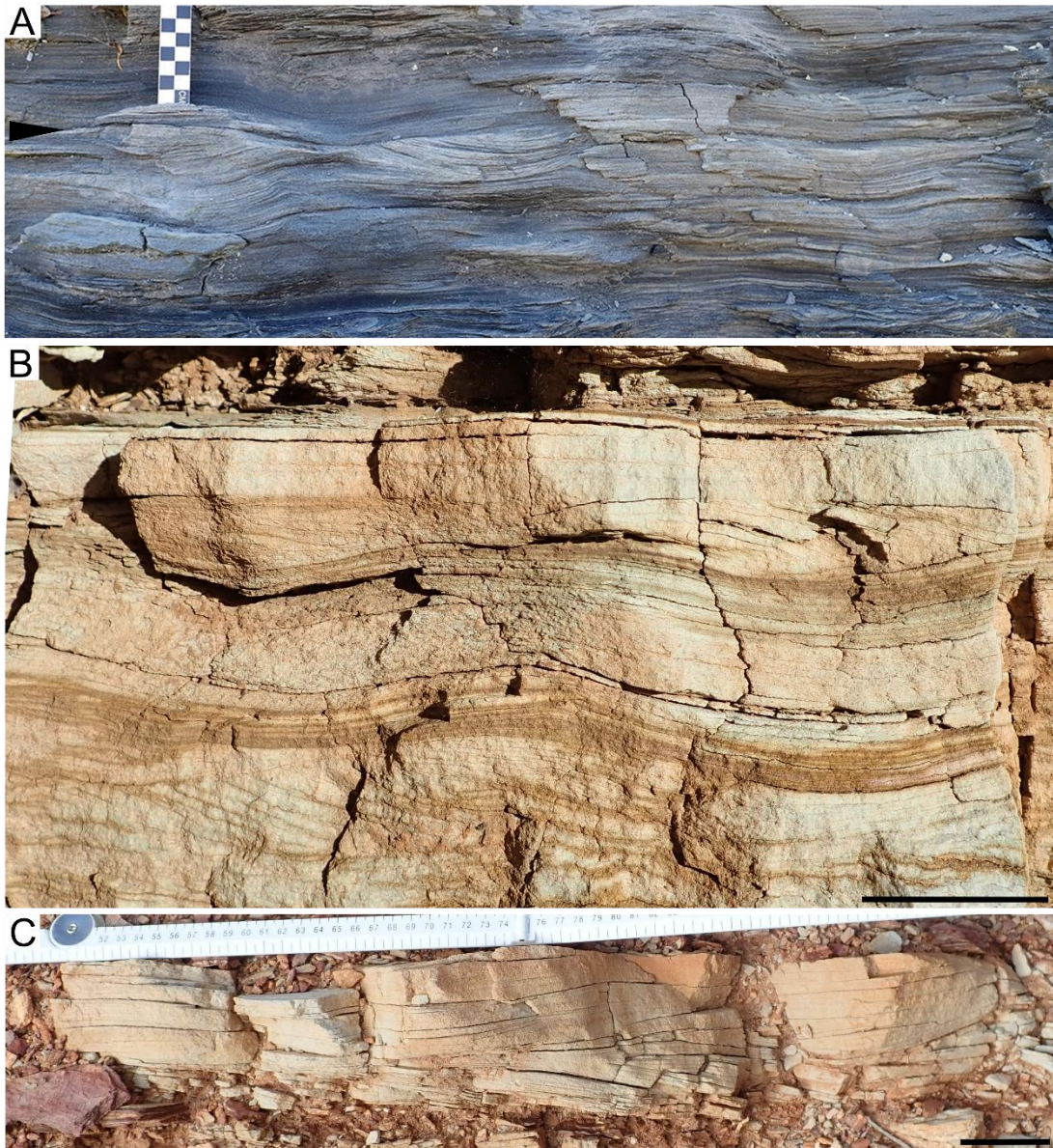


Figure 5.10: Low angle, undulatory cross-stratification from the Witpoort Formation (Facies Su).
A) A heterolithic bed of fine-grained sandstone with mud drapes and black shale interbeds from Waterloo Farm (WF 1b) (black arrow at left indicates a slight erosional discordance that is traceable to a conformable surface). B) Su laminasets with rhythmic mud drapes, interbedded with unidirectional ripple laminae and overlain by horizontally laminated sandstone (Sh) (scale bar = 5 cm). C) An isolated Su interbed in a siltstone horizon (Skitterykloof Member at Waaipoort, WP 3). Scale increments in A are 1 cm, scale bars in B and C are 5 cm. B is from the Rooirand Member (WP 1).



Figure 5.11: Horizontally laminated sandstone (Facies Sh) in the Witpoort Formation. A) Fine-grained, very thinly laminated sandstone topset overlying St (ruler segment is 16 cm long). **B)** Slight mineralogical changes in quartz arenite composition mark the lamination in this photograph. **C)** This lowermost Witpoort Formation bed contains an association of Sh and Sl (p), indicating high energy conditions. **D)** Heterolithic, mud-draped Sh from the upper Weltevrede Formation at Howison's Poort. A plug-shaped burrow is visible at the bottom of the photograph (scale increments in cm). **E)** A fine-grained sandstone interbed from the Rabbit Ridge Upper Section mudstone. This unit blurs the distinction between Su and Sh. A slight erosion surface truncates the sub-horizontal laminae. This horizon contains *Lingulichnus* burrows and deposit feeding traces.

5.2.8. Horizontally-laminated sandstone (Sh)

Description– Sh is characterised by parallel, flat lamination, developed in fine to medium-grained sandstone ranging from light grey or light brown micaceous quartz wacke to light-blue-grey, medium grained quartz arenite (Fig. 5.11 A and B). The facies is pervasive in heterolithic units, but is rare in arenitic successions, where it may easily be confused with current normal exposures of Sp. Horizontal lamination is limited to thin and medium tabular beds (<25 cm thick), normally with clay or silt-draped bounding surfaces. Amalgamated bedsets may be substantially thicker, but this is uncommon in the Witpoort Formation. The facies is more pervasive in the upper Weltevrede Formation where it is strongly heterolithic (Fig. 5.11D). Parting lineations sometimes (uncommonly) ornament the bedding surfaces in both mature, and immature Sh beds. The latter typically has rhythmic muddy partings. Horizontal and undulatory lamination (Su) can be laterally gradational, and appear to form a continuum from flat to low angle (<10°) lamination (Fig. 11E). Trace fossils are recorded in this facies in several places, particularly in the heterolithic beds at the sandstone–mudrock interface (Fig. 5.11D). They include vertical equilibrium traces, deposit feeding traces and record various infaunal activities, but it appears as though surface activity was limited or not preserved in this facies (Chapter 6).

Interpretation– Horizontally- or flat-laminated sand can occur in both upper and lower flow regime current; they may be distinguished by grain size. Lower plane beds occur in sediments with a mean grain size of very coarse sand and greater (Bridge and Demicco 2008; Fielding 2006) and are not considered relevant to facies Sh in the study area. Parting lineation, where it occurs, forms by flow-parallel vortices generated by high-energy supercritical flows, often in very shallow water (Allen 1984; Fielding 2006). Upper plane beds that comprise Sh are formed by very low-relief bed waves that migrate down current (Cheel 1990). Suspended sediment is an important component in the formation of these

bedforms (Van den Berg and Van Gelder 1991; Bridge and Demicco 2008; Ohata *et al.* 2022). High concentrations of fine-grained sediment can suppress the formation of dunes, favouring flat beds with substantially lower current velocities (Best and Fielding 2019). A modern example of this comes from the Yellow River in China, where very high suspended sediment concentrations result in the predominance of flat-lamination (Ma *et al.* 2017). Effects of suspended fine-grained material on the quintessential bedform phase diagrams are increasingly well understood (Van den Berg and Van Gelder 1991; Baas *et al.* 2016; Best and Fielding 2019). Experimental evidence indicates that unidirectional flows may generate parallel-laminated heterolithic sand and mud (Baas *et al.* 2016), contradicting earlier assertions that such facies indicate alternating current and slackwater. Horizontally-laminated sand has likely formed under relatively high current energy with elevated suspended sediment concentrations. Its association with Su implies a related set of causes.

5.2.9. *Ripple cross-laminated sandstone (Sr)*

Description— Sr is expressed in a variety of millimetre to centimetre scale cross-laminated bedforms in the Witpoort Formation. Ripple cross-lamination occurs in fine to medium-grained quartz arenite, but is more common in immature micaceous quartz wacke. Sr is broadly separated into two groups based on the directionality of cross laminae and ripple geometry which are a) climbing ripple lamination and b) combined flow ripple lamination. Both a) and b) can be represented by flaser and lenticular varieties.

The first group, designated Sr (a) includes climbing ripple cross-lamination, which comprises unidirectional trains of cross-stratified laminasets, 0.5–3 cm thick, within beds of up to a metre thick. Sr (a) may be expressed as linear ripple trains with no preservation of ripple crests, indicating a low (subcritical) angle of ripple climb (Fig. 5.12A). Increasing the angle of climb produces crestline preservation, that is termed supercritical climbing ripple lamination (Fig. 5.12B). Straight-crested ripples tend to occur in tabular laminasets whereas

sinuous crested ripples form lenticular or irregular laminasets. A range of ripple forms observed in the Witpoort Formation include a spectrum between these end members. Ripple form sets are occasionally preserved on the tops of cross-stratified beds, they include straight crested ripples, sinuous crested ripples and linguoid ripples (Fig. 5.12G and H). Rib and furrow structures, representing truncated ripple form sets are also common in certain successions.

A second group of ripple forms, Sr (b), is separated from Sr (a) by the multidirectionality of foreset orientation, ubiquity of mud drapes and a greater degree of climb, sometimes tending to aggradation (vertical stacking). This sub-facies regularly preserves ripple form sets, which may be sharp or round crested, and the ripple profile has a greater symmetry than the unidirectional forms described above. Successive laminasets in places show changes in progradation direction creating a herringbone effect (Fig. 5.12D–F). Ripple index (RI) (length to height ratio) was measured at between 5 and 14. These values should be regarded as maxima because any profile that is not crest-perpendicular will increase the apparent RI of the ripple form. Sr (b) occurs in heterolithic successions, interbedded with supercritical Sr (a), Su, Sl and Fh in the study area. Care is required not to confuse Sr (b) with current normal views of sinuous crested Sr (a), which due to the trough shape of the ripple faces can superficially appear symmetrical with bidirectional laminae (Fig. 5.12C). They are distinguished by only containing upwards-concave laminae, whereas Sr (b) show upwards-concave and convex laminae, as well as sigmoidal foreset curvature (Fig. 5.12D and E). Sr (b) is, furthermore, separated from Sr (a) in that the flow vectors are polymodal, whereas the latter have a bimodal directionality with two modes, both within the SW quadrant (Table 5-1).

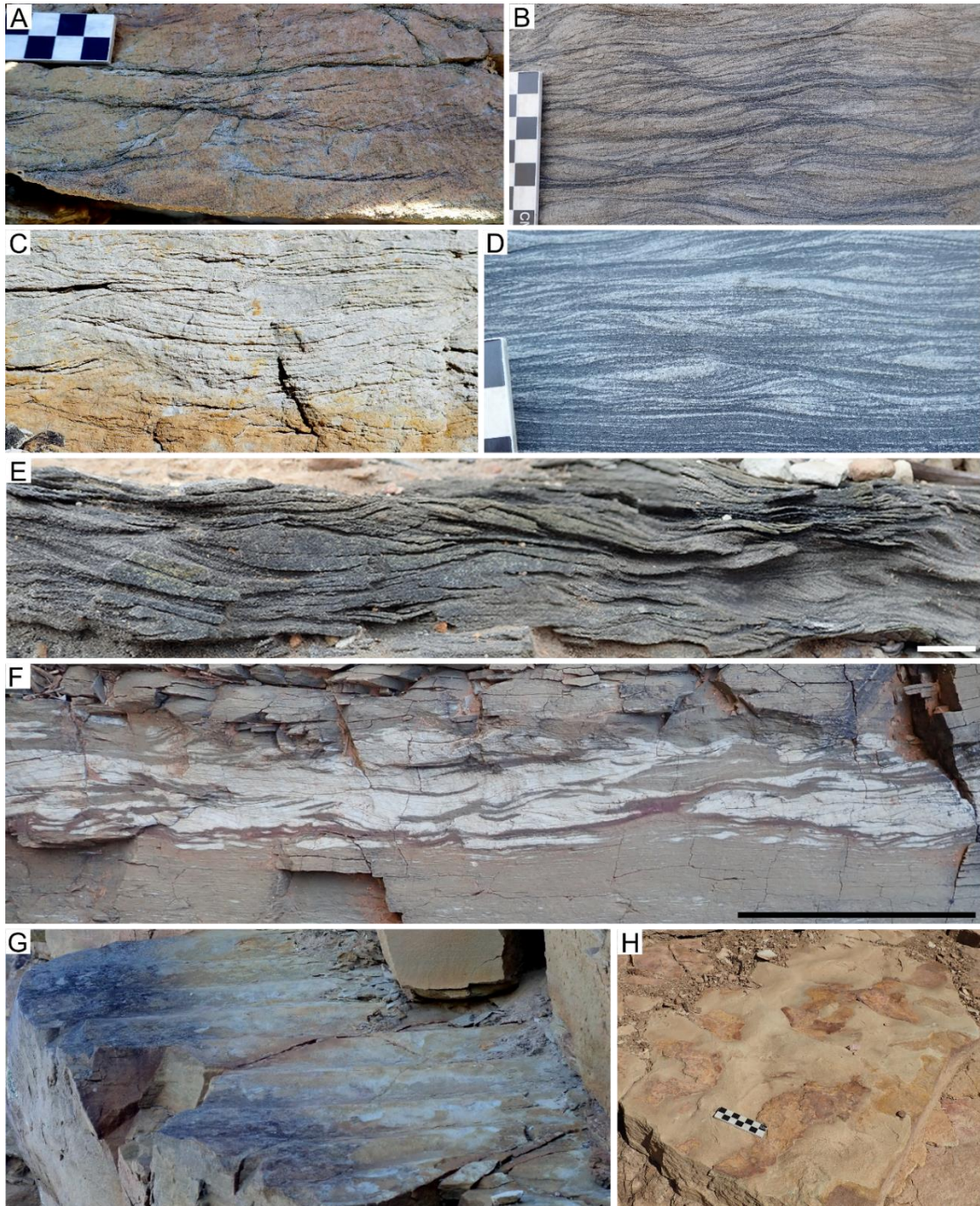


Figure 5.12: Ripple cross-laminated sandstone facies in the Witpoort Formation (Sr (a) A–C, Sr (b) D–F)) (scale bars and squares are 1 cm, unless otherwise stated). A) Unidirectional, ripple cross laminae with curved foresets arranged in laminasets with slight muddy partings. B) Flaser bedding with mud draped ripple cross laminae and laminasets. C) Curved crested, unidirectional ripple cross laminae in current normal view. D) Lenticular bedding with isolated sand ripples embedded in shale, note the bidirectional cross laminae giving a slight herringbone effect. E) Flaser bedding with supercritical climbing ripples. F) Thin bed of round crested, symmetrical ripples with bidirectional foreset orientations and mud drapes (scale bar is 10 cm long). G) Straight-crested current ripple form sets. H) Linguoid ripple form sets.

Interpretation– Ripple cross-lamination can arise in almost any sedimentary environment (Allen 1973). According to numerous lab and field experiments in shallow water, Sr is developed under the same conditions as dunes but with lower flow velocities (Reineck and Singh 1980; Bridge and Demicco 2008). Ripples can only form in sediment grain sizes of coarse sand and smaller particles (Mazumder 2003; Bridge and Demicco 2008). Climbing ripple cross-lamination develops under low velocity unidirectional flows (Walker 1963; Miall 2006). The angle of ripple climb is a function of the rate of ripple progradation versus rate of sediment aggradation. With lower rates of aggradation, the stoss slopes of ripples are eroded, producing sub-critical climbing ripples, whereas higher rates of aggradation produce supercritical climbing ripples with preservation of crests (Ashley *et al.* 1982; Collinson and Mountney 2019). Supercritical climbing ripples tend to occur where an abundance of sediment is carried in suspension (Miall 2006) and this is reflected in the present study area (cf. Fig. 5.12E). Variations in current ripple morphology can be explained by duration of exposure to current. Baas (1999) demonstrated experimentally that, all other variables being equal, ripple forms develop in a sequence from straight-crested to sinuous-crested and finally linguoid ripples with exposure time. Baas therefore argued that linguoid ripples represent the only equilibrium state: the higher the current velocity the more rapid is the progression through the stages. The time required to produce linguoid ripples ranged from minutes to weeks depending on the flow velocity (Baas 1999).

Characteristic features of wave-generated ripples are documented by de Raaf *et al.* (1977). Sr (b) in places exhibits many of these features, including symmetrical profiles, development in pinch and swell sets, bidirectional cross-laminae, aggradational (upwards building) crestlines, draping of crestlines and association with curved, parallel lamination (Fig. 5.12 D–F). Ripple index is a measure that has long been applied for the differentiation of wave from

current ripples and Sr (b) has a RI occupying the overlap area between the higher values known for current ripples and the lower values of wave ripples. This is probably because most of the ripple forms recorded in Sr (b) are the result of combined flows rather than pure current or oscillatory motion. This is indicated by the commonly rounded crests and unidirectional laminasets (Yang *et al.* 2006; Perillo *et al.* 2014).

5.2.10. *Horizontally-laminated shale and mudstone (Fh)*

Description– Fine-grained sediments are concentrated in discrete horizons from a few centimetres up to 12 m thick in the study area (Fig. 5.13A), comprising a small proportion (1–15%) of the overall rock mass in any given section. These horizons contain horizontally-laminated, well sorted silt to clay sized particles, ranging from very thin laminated black shale/siltstone to very thick laminated mudstone. In the eastern sector the facies tends to form more homogenous mudstone and shale beds, whereas further west, sections at Seven Fountains and Waaipoort have more heterolithic expressions of the facies with thinly interlaminated fine-grained sandstone, mudstone and siltstone. Laminated mudrock horizons may be laterally extensive (kilometres in extent) or can occur in restricted lenses. Even where extensive, thickness may vary markedly along strike. These beds may have sharp or gradational lower bounding surfaces, in places grading normally from underlying arenaceous strata, however Fh is almost always sharply overlain by arenaceous facies or conglomerate. Commonly, Fh units are directly underlain by thin conglomeratic horizons (Facies Gi), especially where they exhibit a sharp transition from underlying sandstone.

Repetitive colour stratification occurs frequently in this facies. For example, parallel red oxide banding is present within metre to several metre thick shale horizons in the Howison's Poort, Coombs Hill and Waterloo Farm stratigraphic sections but is absent from other horizons at Coombs Hill and the Rabbit Ridge mudstone horizon. Oxidised laminae are thin (rarely medium) laminated and subordinate in thickness to intervening medium to thick grey

laminae (Fig. 5.13B and E). No variation in grain size was detected between the alternating bands. Hard, ovalular ferruginous concretions are common in association with the red oxide layers.

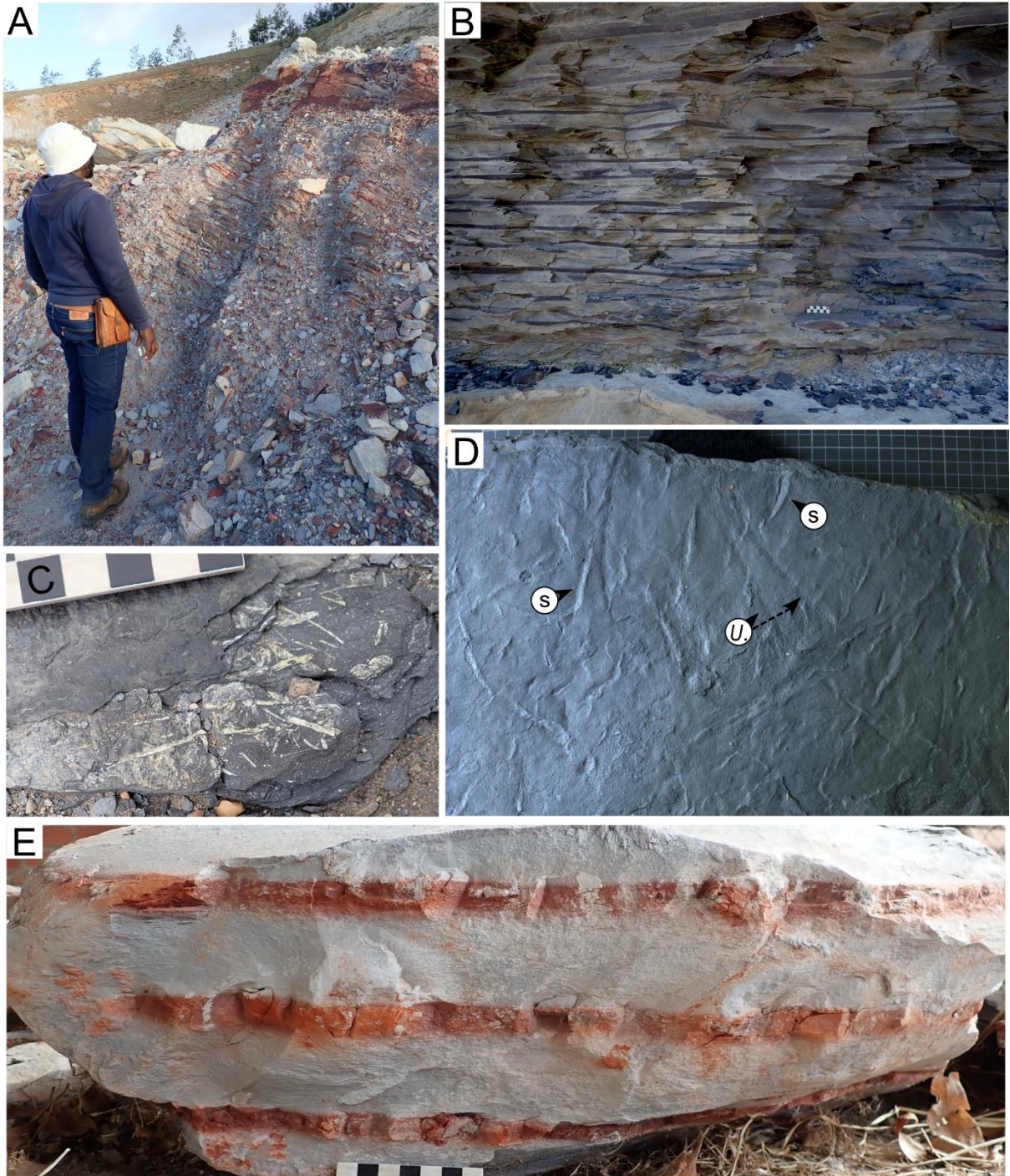


Figure 5.13: Horizontally-laminated shale and mudstone facies (Fh). A) A 3-metre thick mudstone horizon showing regular colour stratification (Coombs Hill, IO horizon). B) A metre thick black

shale horizon with regular thin oxide rich interbeds from Howison's Poort (HP 3). C) Finely preserved comminuted plant fragments in black shale at the Waterloo Farm Blue Layer (scale in cm). D) Synaeresis cracks associated with *Undichna* from the Waterloo Farm Main Fish Lens. E) Block of mudstone from the horizon shown in A. Stratification in this rock is limited to horizontal alternating red and grey laminae. The red, oxidised layers are discontinuous, and are the product of secondary iron oxidation. They are subordinate in thickness to the grey layers.

Plant fossils in various states of fragmentation are a nearly ubiquitous feature of this facies, ranging from nearly complete plants to miniscule fragments. Certain layers contain mats of comminuted terrestrial plant matter and/or algae. Plant remains within such mats often show sub-parallel alignments (Fig. 5.13C). This plant matter (n=3) was aligned along NE–SW to NNW–SSE trends. Invertebrate body fossils are less widespread, only being known from five separate shale horizons, three of which occur in the stratigraphically poorly constrained Coombs Hill section (Harris *et al.* 2021a). Invertebrate taxa from Fh in the Witpoort include lingulid brachiopods, bivalve molluscs, putative ostracods, eurypterids and scorpions. Trace fossils in this facies are common, but poorly preserved, usually diminutive invertebrate burrows or vertebrate coprolites. Fish swimming traces are recorded at one stratigraphic horizon. Many horizons of this facies are, however, apparently unbioturbated.

Interpretation– Depositional rate of muds varies over several orders of magnitude depending on the depositional setting. Deep-marine muds can accumulate as slowly as millimetres per thousand years (da Silva *et al.* 2020), whereas estuarine and lagoonal muds can attain depositional rates of tens of centimetres per year (Shephard and Moore 1960; Reineck and Singh 1980; Cowan *et al.* 1998), up to metres per year in inner estuarine deposits close to the fluvial tidal transition (Dalrymple *et al.* 1992). Deposition of mud on modern tidal flats has been recorded at a rate of metres per year (Hantzschel 1939), and in ancient tidal flats up to 3.8 m per year (Lanier *et al.* 1993). Palaeontological analyses indicate lagoonal (Hiller and Taylor 1992; Harris and Gess 2022 (this work)) and estuarine (Gess and Whitfield 2020) provenances for shale and mudstone in the Witpoort Formation.

Horizontally laminated muds are widely considered to represent suspension fall out in still water settings (Boggs 2006; Nichols 2009). Recent work has, however, shown that muds can accrete under rapid currents (up to 2m/s), particularly where suspended sediment concentrations are high (Baas and Best 2002; Dalrymple and Choi 2007; Mackay and Dalrymple 2011; Schieber 2011). Fluid mud, which is a viscous, mobile slurry of bottomwater mud can occur where suspended sediment concentration exceeds 10g/l (Mackay and Dalrymple 2011). This is particularly important in estuaries due to the trapping of suspended sediment, which may far exceed this threshold and may vary by orders of magnitude within the system (Uncles *et al.* 2006). Suspension fallout and fluid mud deposition operate simultaneously within such systems (Dalrymple and Mackay 2011), and this is a plausible explanation for the thicker laminated mudstones with rapidly interred plants (further explored below, FA 2c). This analysis should be complimented by a detailed microfacies analysis of the variations in mudrock lamination, which can potentially differentiate dynamic from suspension settled muds (e.g. Mackay and Dalrymple 2011, Schieber 2011). It is possible that both suspension fallout and dynamic mud deposition account for Fh accumulations in the Witpoort Formation, processes that are not mutually exclusive.

Dropstones

As an addendum to Fh, there exist in the collections from Waterloo Farm anomalous sand clasts, the like of which were not encountered in the studied sections. Three pebbles of fine-grained quartz wacke are curated in the Albany Museum collection. They comprise fine-grained quartz wacke embedded in black shale from the Waterloo Farm Main Fish Lens (MFL). The clasts are rounded medium to coarse pebbles, approximately oval in profile, with slightly tapered margins, and coarser grained than the black shale host rock. The latter point excludes the possibility that the clasts are concretions. None of the clasts are associated with

layers of coarser-grained material, and it appears as though mud deposition continued without interruption before, during and after the clasts were laid down. One of them, which was collected *in situ* by R.W. Gess is shown in Figure 5.14. It is eye-shaped in profile with lateral margins that pinch out slightly, and the black shale laminae below the clast are deformed downwards. Above the pinch-out margins the shale laminae onlap the clast. The overlying laminae are deformed upwards. Deformation of the laminae may be due to the coarser grain of the clast relative to the matrix, and its resulting resistance to compaction. Impact of the clast dropping from above may also have deformed the underlying laminae to some degree. Deformation at the lateral margins of the clast itself indicates that the clast was soft at the time of burial.

The likely transport mechanism for the clasts is that they were rafted and dropped onto the substrate from above. Three mechanisms for the rafting of dropstones discussed by Doublet and Garcia (2004) include stone swallowing by vertebrates and rafting by floating vegetation or by ice. Only the latter two are considered plausible explanations for these dropstones. Floating vegetation must have been abundant within the MFL's locus of deposition, however this mode of transport does not explain the rounded shapes of what were clearly semi-consolidated intraclasts, or their prior consolidation. The simplest explanation therefore is that they were carried in a frozen state by floating ice. Melting after impact with the substrate enabled the slight deformation seen along the clast margins.



Figure 5.14: A dropstone from Waterloo Farm contained within a block of weathered black shale. Note the slightly tapered margins of the clast, interpreted to be the surface level that the clast occupied after impact where the margins subsided slightly. Above this level shale laminae onlap the clast.

5.2.11. *Rhythmic laminated sandstone with mud drapes (Sc)*

Description– This facies is a subset of several sandstone facies described above, and is distinguished based on the presence of mud drapes which are often coupled. Irregular, sometimes rhythmic, thickness variations in occur in the sandstone laminae. Facies Sc most commonly comprises horizontally laminated sandstone, but is also recognised in trough cross-stratified, planar cross-stratified, low-angle cross-stratified, undulatory and ripple cross-laminated sandstone. It is best developed in the upper Weltevrede Formation where it forms parts of rhythmically laminated heterolithic beds up to a metre thick. In the Witpoort Formation the facies occurs in thin to medium bed units, typically developed in units immediately underlying (or sometimes overlying) laminated shale and mudstone (Fh).

Mud couplets are here referred to in the sense of Visser (1980), which denotes a pair of parallel, muddy or silty laminae separated by a sandy lamina. They are often amalgamated into a single mud drape when traced laterally. These drapes are very thin to thin, but may form thicker muddy accumulations, particularly in the toe-sets of cross-strata. Commonly sand laminae occur in laminasets that show cyclical thinning and thickening, which is most recognisable in horizontally laminated forms (Fig. 5.15A–E). One example from the strata immediately underlying a thick mudstone unit at Coombs Hill is a medium bed of horizontally laminated fine-grained sandstone with gutter casts at the base. It contains between seven and 11 mud draped sand laminae in a set (Fig. 5.15A and B). Examples of similar facies from Waterloo Farm (WF 2) indicate between four and nine laminae in a set. In these examples the laminasets are irregular in thickness and number of laminae, which vary markedly within a bed. The presence of mud drapes and couplets is also sporadic. Strata from the upper Weltevrede Formation at Waaipoort (Fig. 5.15C) have a more consistent rhythmicity in the laminasets. The precise number was not quantified during this study, but appears to contain a significantly greater number in each set.

Cross-stratified sandstone with mud-draped foresets attain a variety of scales from thin bedded to thick bedded. Usually foresetting is irregular, with no clearly repeated cyclicity, but laminae are highly variable in thickness. Reactivation surfaces are common, separating laminasets that are muddier in the direction of forward accretion (Fig. 5.15G). Mud drapes are sporadically preserved on foresets but tend to form much greater accumulations in the toe sets. Often the foresets are sigma-shaped with a more or less horizontal toe set and top set attached to an inclined foreset (Fig. 5.15F). The defining features of this facies are; irregular foreset thickness, presence of mud drapes and couplets, rhythmic laminasets, and, where cross-stratified; sigma-shaped foresets and reactivation surfaces.

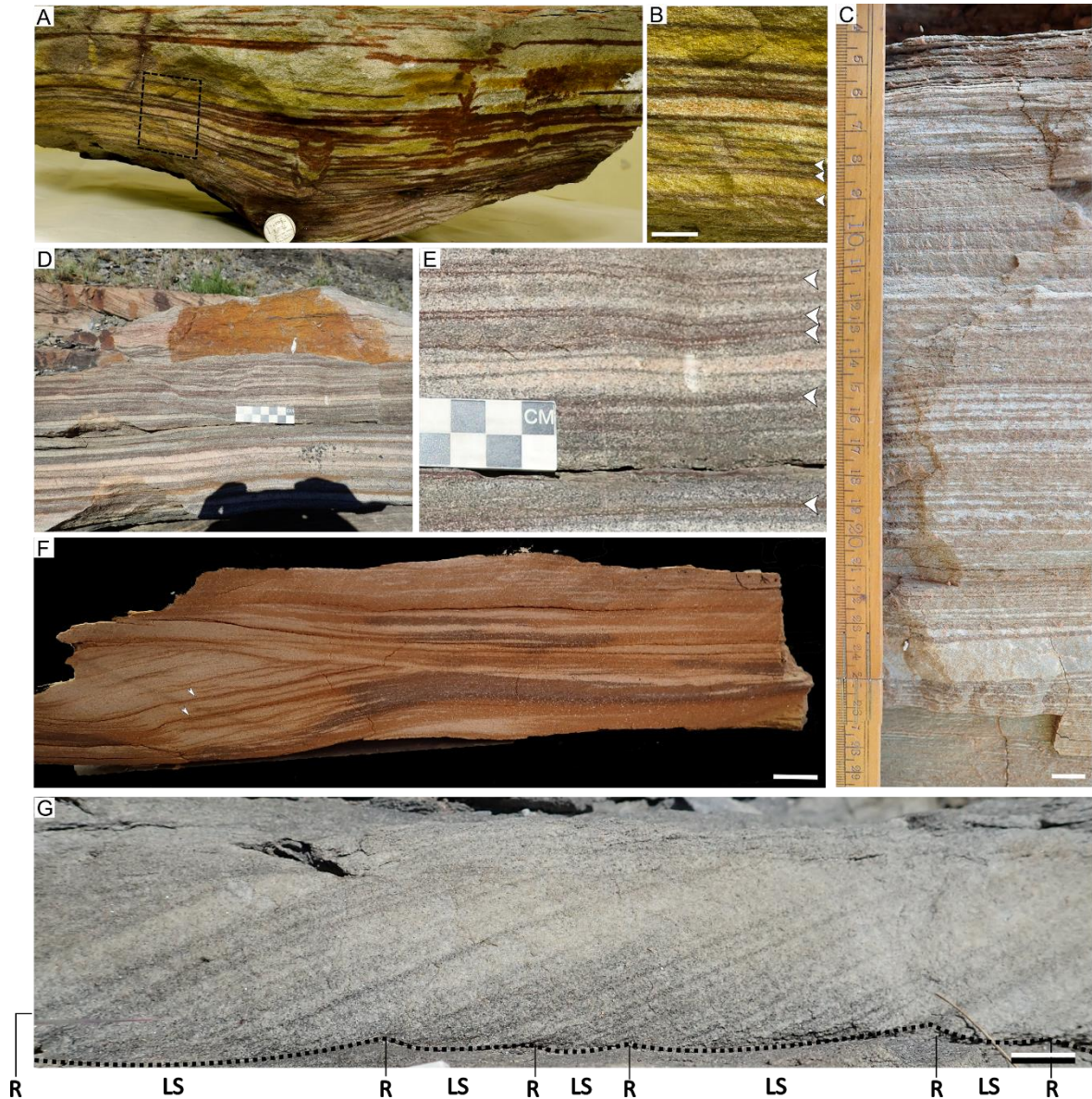


Figure 5.15: Sandstone with mud drapes facies (Sc) (All scale bars and squares represent 1 cm, white arrows indicate mud couplets). **A)** A rock sample from Coombs Hill containing irregular parallel-laminated rhythmites. A gutter cast at the base of the slab is concordantly filled. Trace fossils are *Skolithos* ichnofacies taxa with a very low bioturbation intensity (BI 1–2). Stippled rectangle indicates the area enlarged in B, which shows the mud drapes and couplets. **C)** A 25 cm thick bed (uppermost Weltevrede Formation, Waaipoort) containing horizontally-laminated sand with mud drapes, arranged in 1–5 cm thick laminasets of cyclically fining and coarsening laminae, as interpreted to the right of the photograph. **D)** A bed of subhorizontal, parallel-laminated quartz arenite with sporadic very thin mud drapes and couplets, which occupies the unit immediately underlying a black shale bed, the Waterloo Farm upper shale (seen at top of photograph). **E)** Enlargement of the bed in D showing the mud couplets (arrowed) and heavy mineral laminae. **F)** Sigmoidal cross-strata with horizontal, mud rich topset and bottomset (upper Weltevrede Formation, Howison's Poort). Many of the mud drapes are confined to the bottomset, some of them continue along the length of the foreset. The most continuous drapes are distinctly coupled

(arrows). **Multiple slight reactivation surfaces are evident in the bed. G) Planar cross-stratified thin bed with irregular foreset thickness and alternating silt-poor and silt-rich laminae. The laminasets (LS) are separated by slight reactivation surfaces (R) marked by undulations at the lower bounding surface (stippled line) overlain by the most sand-rich laminae. A horizontal topset erosively overlies the bed.**

Interpretation– Mud-draped sand laminae may form in many non-marine and marine depositional environments. Paired mud drapes are, however, considered to be distinctive of tidal depositional settings (Visser 1980; Tessier and Gigot 1989; Nio and Yang 1991; Dalrymple *et al.* 1992; Fenies *et al.* 1999; Hovikoski *et al.* 2005). The mud couplets are thought to represent fallout from suspension during tidal slackwater (Visser 1980; Hovikoski *et al.* 2005; Mackay and Dalrymple 2011). Tidal couplets may arise in two ways, one, by deposition during a dominant tide and its subsequent slackwater forming a thick lamina and drape, followed by the subordinate tide and slackwater forming a thinner lamina and drape (Visser 1980; Kuecher *et al.* 1990). A more common cause for the generation of tidal couplets is diurnal tidal inequality, which is the unequal magnitude of successive semi-daily tides (Kvale *et al.* 1995; Archer *et al.* 1995; Choi and Park 2000). In this case, only the dominant tide (flood or ebb) is depositional, and will therefore produce only half the number of laminae in a neap-spring cycle. The preservation of tidal signals is favoured in macrotidal depositional settings, whereas microtidal settings are unlikely to produce complete tidal rhythmites (Dalrymple *et al.* 1992). Specific microenvironments within the intertidal zone, such as abandoned channels, migrating tidal point bars, tidal deltas and mouth bars have the greatest potential for preservation of tidal rhythmites (Choi and Park 2000). Even in successions where fully developed tidal rhythmites occur, they can be rare, as for example in the Pennsylvanian Francis Creek Shale, where only 10% of laminated tidalites exhibited fully developed tidal rhythmicity, whereas the remaining 90% was irregularly laminated (Kuecher *et al.* 1990; Feldman *et al.* 1993).

Rock samples from the Witpoort Formation on which lamina counting was performed show cyclicity with the number of events in each cycle being far below the number expected for typical tidal bundles. Incompleteness in the tidal depositional record can be the result of many factors (e.g. Boersma and Terwindt 1981; Roep 1991; Dalrymple *et al.* 1992; Cowan *et al.* 1998; Mackay and Dalrymple 2011). The highest energy parts of the tidal cycle (spring tide) may be erosional and may produce only a reactivation surface, whereas the lowest tidal energy during neap tides may be insufficient to produce a deposit in some parts of the intertidal zone (Dalrymple *et al.* 1992). Removal or nondeposition of parts of the tidal cycle can be explained by the persistent growth and decay of sandbars, the interplay between tidal energy and river discharge, or the influence of storms (Boersma and Terwindt 1981; Roep 1991). In the fluvially controlled parts of estuaries, irregular tidal signals can be generated by the modulation of fluvial currents, where the largest (spring) tides are the most likely to form mud drapes (Mackay and Dalrymple 2011). The described deposits exhibit many features that are typical for tidal deposits including sigmoidal foresets, irregularity in foreset thicknesses, thick-thin cyclicity, mud drapes and couplets. The putative tidalites described above may therefore represent incomplete tidal bundles, which are discussed in their stratigraphic and palaeo-environmental context below.

5.2.12. *Soft sediment deformation structures (SSDS)*

Description— SSDS are common in the Witpoort Formation, increasing in frequency towards the top of the formation. They include parabolic cross-stratification, flame and chevron structures, wrinkle marks, convolution, slump structures, load casts, syn-sedimentary faults, bedding-parallel grooves and a clastic dike. Parabolic (‘recumbent’ or ‘oversteepened’) cross-stratification is almost ubiquitous in deformed sandstone horizons, sometimes associated with water escape structures and commonly accompanied by convolutions. Mud-rich strata are affected by small-scale convolutions and synaeresis cracks,

whereas the mudstone–sandstone interface shows a different style of deformation, with slumps, load casts and mud flames.

Parabolic cross-stratification

Parabolic cross-stratification affects cross-stratified sandstone facies, where the foresets are steepened beyond the angle of repose or overfolded. Parabolic beds are 5 cm to 40 cm thick, and can occur at multiple horizons within a bedset separated by beds that are not deformed (Fig. 5.16). Small-scale convolutions are common in the strata overlying parabolic beds (Fig. 5.17D and G). Parabolic folding is symmetrical or asymmetrical. Foresets can be recurved at about their midpoint so that the upper part mirrors the lower part with an axis of deformation that is parallel to the lower bounding surface (Fig 5.17A). The axis of deformation can also occur lower or higher along the foresets and is sometimes oblique to the lower bounding surface (Fig. 5.17F and G). Substantial thickening of foresets occurs along the axis of folding and they have usually been truncated at the upper bounding surface, but may turn sub-horizontal at the topset. This is termed simple parabolic cross-stratification. Simple parabolic deformation is common, but at several horizons more complex forms occur. This includes foresets deformed in an S-shape with two reversals of foreset orientation, the tops of foresets being returned to an orientation similar to that observed at the base. Multiple reversals of foreset orientation can occur within a foreset, resembling convolution (Fig. 5.17B). Parabolic horizons can extend for tens of metres laterally, although the horizons may be laterally restricted when they pinch out or are truncated. Their maximum extent is hard to gauge from present evidence.

Parabolic cross-stratification is related to soft-sediment deformation at the sediment–water interface by unidirectional current drag following liquefaction of sediments, possibly resulting from seismic shock (Allen and Banks 1972; Bhattacharya and Saha 2020), although rapid changes in flow velocity have also been suggested (Røe and Hermansen 2006;

Whitmore *et al.* 2015). The latter is supported by experimental evidence; Hendry and Stauffer (1975) described parabolic cross-stratification from Pleistocene periglacial sandstone in Canada. They suggested, based on laboratory experiments (McKee *et al.* 1962; Middleton 1967), that hypersaturated sand-rich flows caused the dragging. McKee *et al.* (1962) formed parabolic cross-stratification by two methods; using a high-pressure hose and by dragging a sand bag across a sediment surface. Parabolic cross-stratification is considered to be most common in fluvial environments (Allen and Banks 1972; Røe and Hermansen 2006), but has also been reported in subtidal (Mazumder and Altermann 2007) and shallow-marine strata (Whitmore *et al.* 2015). Convolutions in the overlying horizons are explained by post-depositional liquefaction of subsequent deposits because the buried deformed horizon contains a resettled, and therefore more compact zone, forming a density contrast in the substrate (Owen and Moretti 2011).



Figure 5.16: Repeated parabolic cross-stratification alternating with undeformed thin beds in the upper Witpoort Formation (Ruler is 50 cm²).

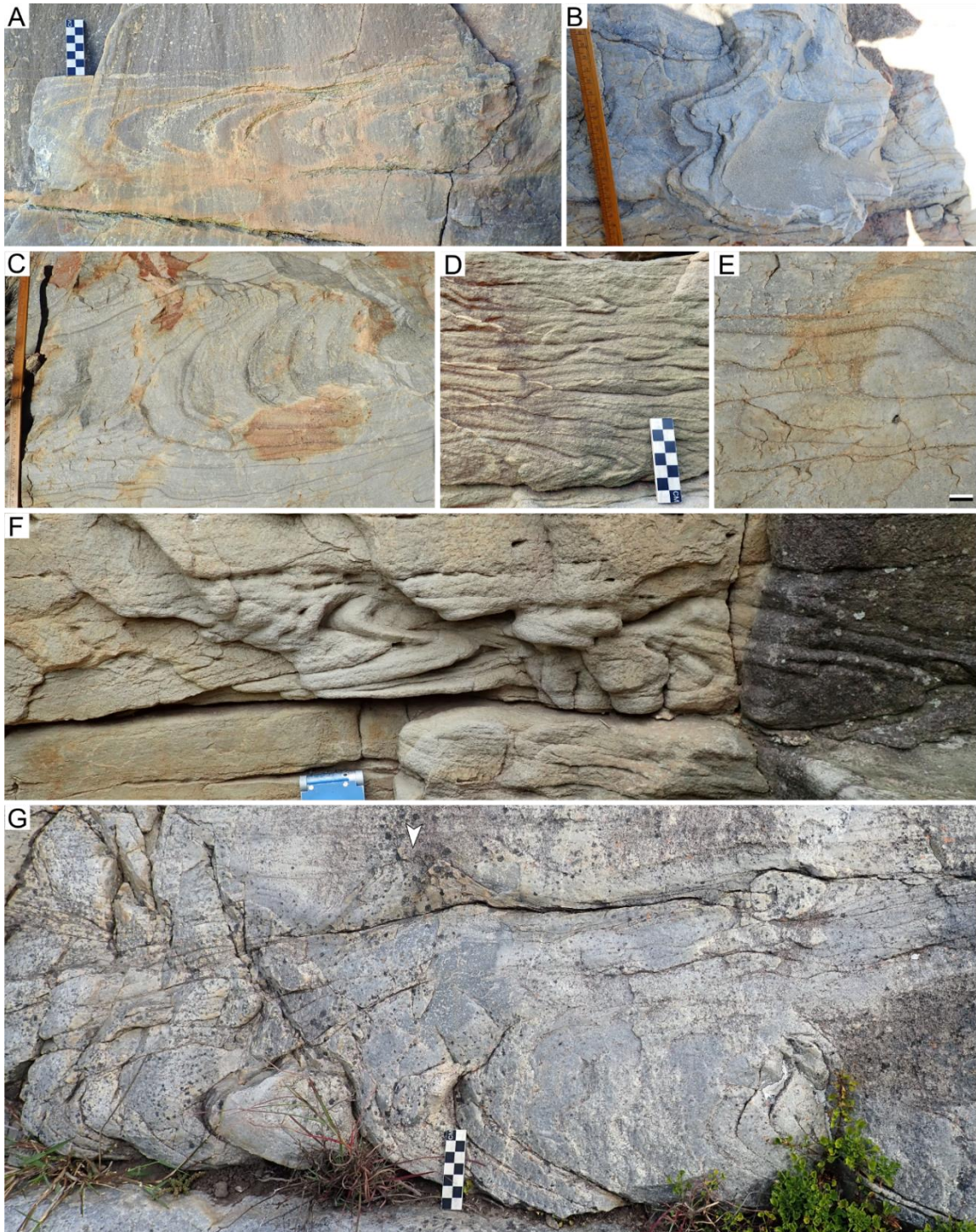


Figure 5.17: Parabolic cross-stratification in the Witpoort Formation: A) Simple, symmetrical parabolic cross-stratification interbedded with undeformed strata in the lower Witpoort Formation at Howison's Poort (scale in cm). B) Convoluted parabolic cross-stratification from the upper Witpoort Formation in the Grey Dam section (visible ruler segment is approximately 20 cm long). C) The axis of deformation in this set is oblique to the set lower bounding surface (ruler segment is approximately 35 cm long). D) Small-scale parabolic cross-stratification overlain by thinly laminated sandstone with abundant small-scale convolutions in the uppermost Witpoort Formation strata at a quarry adjacent to Grey Dam (scale in cm). E) Enlargement of a flame in

sandstone immediately underlying the deformation shown in C (scale bar represents 1 cm). F) Foresets are deformed into an S-shape (compass is 72 mm in diameter). G) Simple parabolic grading laterally to S-shaped parabolic cross-stratification in the lower Witpoort Formation at Featherstone Kloof. Note convolutions and dewatering structures in the overlying horizon (arrow indicates a truncated chevron structure) (scale in cm).

Syn-sedimentary faults

Syn-sedimentary faults are observed in one place (WF 2) developed in horizontally-laminated sandstone underlying a thick shale bed. This horizon overlies a parabolic cross-stratified set. Deformations include normal faults with millimetre-scale displacements across which sedimentary laminae are bent but not interrupted (Fig. 5.18C).

Chevron structures

Chevrons are upwards-convex deformations in which sedimentary laminae are continuous forming inverted cone-in-cone structures (Fig 5.18A and B). They occur in sandstone facies at a variety of scales with displacements of centimetres to tens of centimetres. Bedding perpendicular and oblique axes of deformation occur. In one place a chevron structure overlying a parabolic cross-stratified horizon is truncated, indicating post-deformation erosion at surface (Fig. 5.17G). Abundant small-scale convolutions can occur at the same horizons and along the chevron walls. Bedding-oblique chevrons are locally developed beneath parabolic cross-stratification in the Grey Dam section (Fig. 5.17C and E), overlain by similarly orientated convolution in the parabolic beds. Here water escape structures interact with parabolic foresets in such a way that suggests their contemporaneity. Chevron structures are interpreted as representing localized zones of water escape (Bhattacharya and Saha 2020).

Load casts

Load casts, indicated by rounded sandstone bodies emplaced into underlying mudstone or siltstone, are recognised at various horizons where sandstone sharply overlies fine-grained strata. They include centimetre and decimetre scale structures (Fig. 5.18 C, D and E). Their

occurrence at this interface implies that they formed through instability in the sediment related to density contrasts (van Loon 2009).



Figure 5.18: SSDS in the Witpoort Formation: A) Bedding-oblique chevron structure from a horizon immediately overlying the structure shown in G (ruler is 50 cm squared). B) Small-scale, bedding-oblique chevrons. C) A syn-sedimentary fault. D) Centimetre-scale load casts at the sandstone-siltstone interface in heterolithic deposits. E) Load casts at the siltstone-sandstone interface (scale bar = 5 cm). F) Complex convolution structures with pipe-like features (arrowed)

interrupting sedimentary laminae (ruler increments in cm). G) Large scale loading and slump structures at the sandstone-shale interface, abundant SSDS are recorded in the immediately overlying sandstone (ruler is 1 m long).

Synaeresis cracks

Synaeresis cracks are abundant on certain black shale bedding surfaces in the Waterloo Farm MFL (Fig. 5.19), and in similar facies elsewhere in the study area. At Waterloo Farm, they are short, linear or sigma-shaped structures, up to 8 cm in length, which are widest in the middle and taper towards both ends. The crack is a narrow structure, less than a millimetre in diameter, but its oblique intersection with the bedding surface gives it an apparent width of up to 1 cm. All the cracks intersecting a bedding surface are similarly orientated at an oblique angle to bedding and sub-parallel to each other. The sediment surface is normally slightly raised on the footwall of the crack. In profile, they penetrate at least 1–2 cm of rock thickness at an angle of about 40° from the horizontal (Fig. 5.19)

The described structures are consistent with subaqueous shrinkage, rather than desiccation, cracks because of their sub-parallel alignments, discontinuity and the linear spindle shape (McMahon *et al.* 2017). Widely held explanations for subaqueous synaeresis cracks are seismic activity (Pratt 1998), salinity changes (Hovikoski *et al.* 2008), or microbial activities (Harazim *et al.* 2013). All these explanations invoke shrinkage due to the expulsion of water from flocculated or swollen clay particles in the substrate (McMahon *et al.* 2017). The exact causes for synaeresis cracks are debated, and it is likely that more than one mechanism for their formation exists. Their occurrence in laminated black shale from Waterloo Farm in the study area is consistent with the conventional explanation of shrinkage due to salinity changes in the water column (Hovikoski *et al.* 2008; van Loon 2009; Collinson and Mountney 2019).

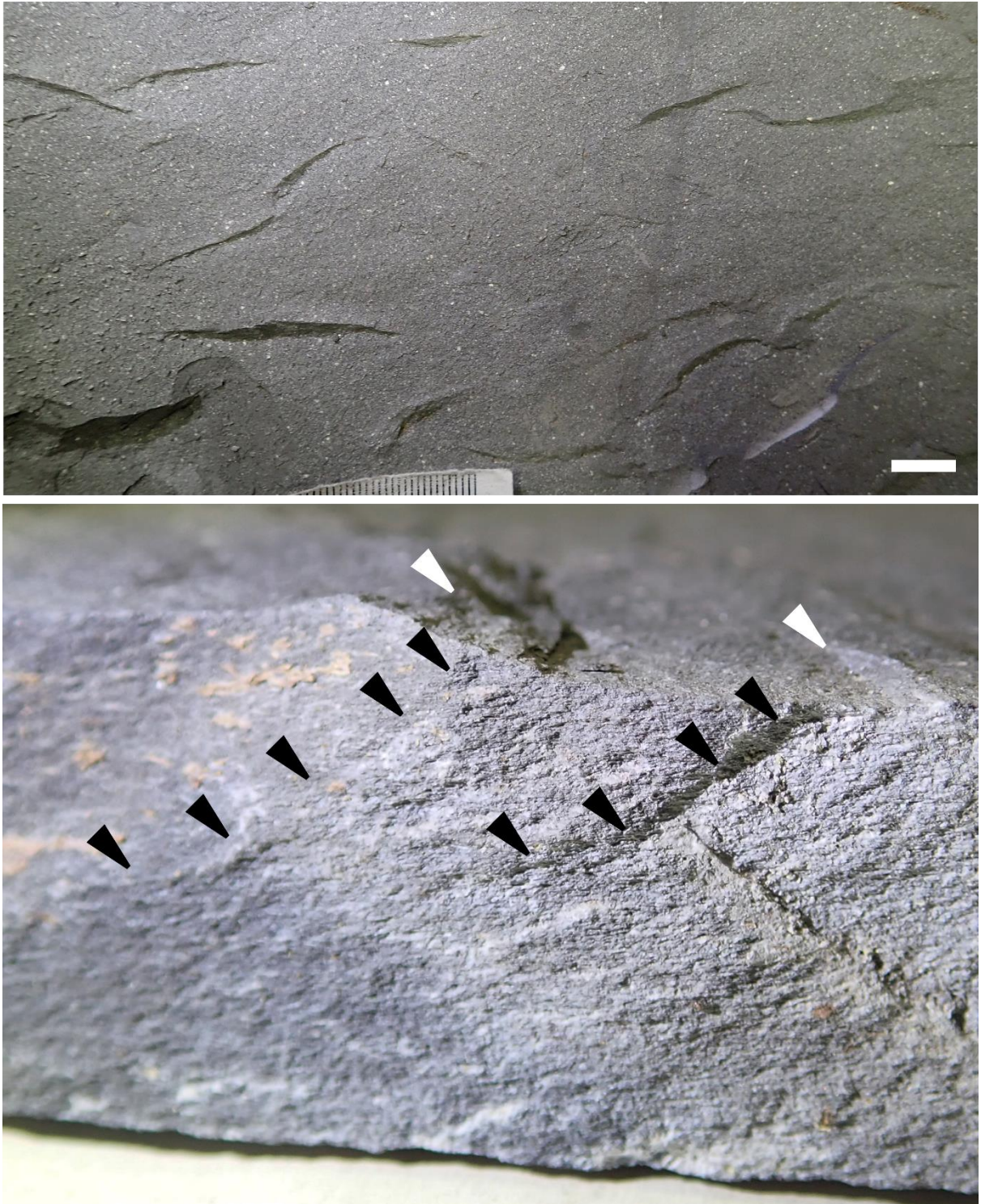


Figure 5.19: Synaeresis cracks from black shale facies (Fh). Above, the bedding plane expression (scale bar = 1 cm). Below, profile view of the same slab as above, showing two parallel cracks that intersect the bedding plane (white arrows), and their intrusion into the underlying bed (highlighted by black arrows).

Clastic dyke

A clastic dyke is exposed in two places in the Waterloo Farm section on opposing sides of the national highway. It is a straight-walled bedding-perpendicular structure filled with structureless sandy, micaceous siltstone. No displacement of marker beds on either side of the dyke walls was detected.

In the north facing cutting (WF 1a) the structure is approximately 1 m wide (Fig. 5.20A). On the opposing cutting (WF 1b) the structure is approximately 1.5 m wide at road level, and widens to 2 m near the top of the section. In both sections the structure cross-cuts the entire exposure and neither its top nor base is located in either section. Above the road cuttings there is no sign of the structure due to soil cover, nor is it visible in satellite imagery. The vertical extent of the structure exposed in WF 1b is roughly 18 m perpendicular to bedding. Lamination within the dyke walls is sharply truncated, and the laminae are not deformed. Wall-parallel silica deposits up to 5 cm thick occur sporadically along the dyke walls (Fig. 5.20C). The dyke fill is homogenous dark grey micaceous sandy siltstone, which is locally weathered to light grey and it contains some ferruginous concretions in places (Fig. 5.20D). No grain size changes were detected in the field, but the matrix becomes progressively lighter in colour and softer in texture towards the top of the cutting, with proximity to the original surface (Fig. 5.20D). The dip changes slightly with steeping of the strata from the southern to the northern cutting, but strike remains similar with a 2° difference between exposures. Back rotation of the structure at both exposures (Fig. 5.20E) shows that was originally orientated sub-perpendicular to bedding (88°) extending WNW to ENE.

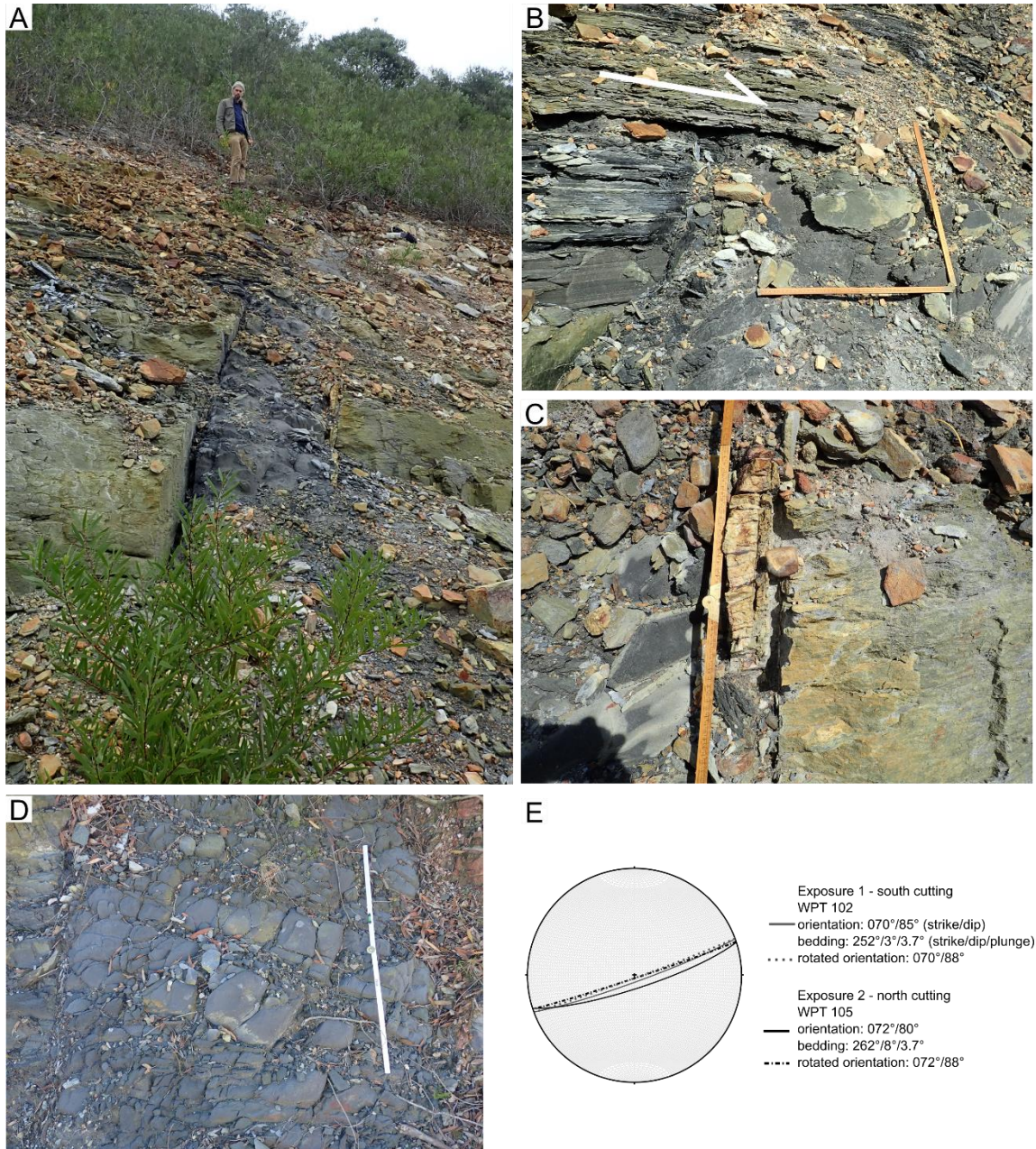


Figure 5.20: A clastic dyke exposed at Waterloo Farm, note the sharply truncated, undeformed bedding in the dyke walls in A–D. A) Profile view of the feature, neither the top nor base of the structure are exposed in the section. B) Bedding-parallel slip related to the regional structural deformation is evidenced by displacement of the dyke walls. Arrow shows the direction of displacement. C) Wall-parallel silica deposits. D) Massive dyke fill. E) Stereonet restoration of the structure shows that it was originally vertical.

Orientation perpendicular to bedding and argillaceous fill indicate an originally sub-vertical, primary sedimentary structure. A fault origin is excluded because no brecciation was detected in the structure or along the walls, nor is there any displacement detected in concert

with the dyke. Furthermore, it does not align with orientations recorded for faults along this exposure. The important feature is that bedding in the dyke walls is undeformed. If the dyke intruded soft sediment, it would be expected that the bedding would have been turned upwards (if the material was injected from below), or to have slumped if the dyke was actively or passively filled from above. The sharp walls suggest that the dyke was either intruded into lithified country rock or a frozen substrate (Larsen and Mangerud 1992). The structureless siltstone filling is consistent with injection, whereas passive fill would tend to produce a laminated fill (Lunina and Gladkov 2016). At present we have not identified the source of the material to conclusively determine whether the dyke has been filled from above or below. The dyke is expected to cross-cut a thick black shale near road level in WF1b. Excavation of the overlying vegetation and soil at this level could ascertain whether the material originated at this horizon, if indeed it had been injected from beneath. Perhaps the most analogous geological structures from the literature are sub-vertical, parallel-sided diamictite-filled dykes from the Falkland Islands, which are determined to have been formed by subglacial injection during the earliest Carboniferous to Permian glaciations (Hyam *et al.* 1997). The Falkland Islands dyke swarms occupy host rocks of similar ages and lithology to that here described, but the main difference is in the absence of wall-parallel lamination or diamictite clasts in the dyke fill from Waterloo Farm.

Bed-parallel grooves (ice keel scour marks?)

Bedding parallel grooves and striations are recorded at two horizons in the study area, both in the upper Rooirand Member underlying the highest prominent shale horizon below the Perdepoort Member. The striae are straight, parallel grooves with flat or curved bottoms and lateral push-up ridges. The grooves resemble ripple troughs, but are much straighter and more variable in size than ripple forms observed in the study area. In Waterloo Farm section 2 the features trend NNE–SSW, and are recognised on one bedding surface in a metre thick

channelised unit that underlies a 5–6 m thick shale bed. Similarly, at Waaipoort, near the top of the Rooirand Member, the grooves occur about 1.5 m below a 3 m thick shale bed, and trend ENE–WSW. These horizons appear to both represent a similar stratigraphic level. At both horizons the grooves are variable in size from less than 1 cm up to 10 cm in diameter. Smaller grooves may be superimposed within larger ones.

Initially the features were labelled ‘pseudo-ripples’ in the field, because they are clearly different from ripple crests and troughs, being much more parallel sided and irregular in diameter. A structural origin was considered, however, the features do not resemble the slickensides that are common in these sections, which normally occur as quartz mineralisations on flat surfaces. Their observation at only one level in two sections counts against a structural origin. The texture of these ‘pseudo-ripples’ is rather related to soft-sediment deformation, as indicated by their observation in profile (Fig. 5.21C). They strongly resemble ice keel scour marks, which are formed by the dragging of floating ice over frozen, unconsolidated sediment (Veseley *et al.* 2015).

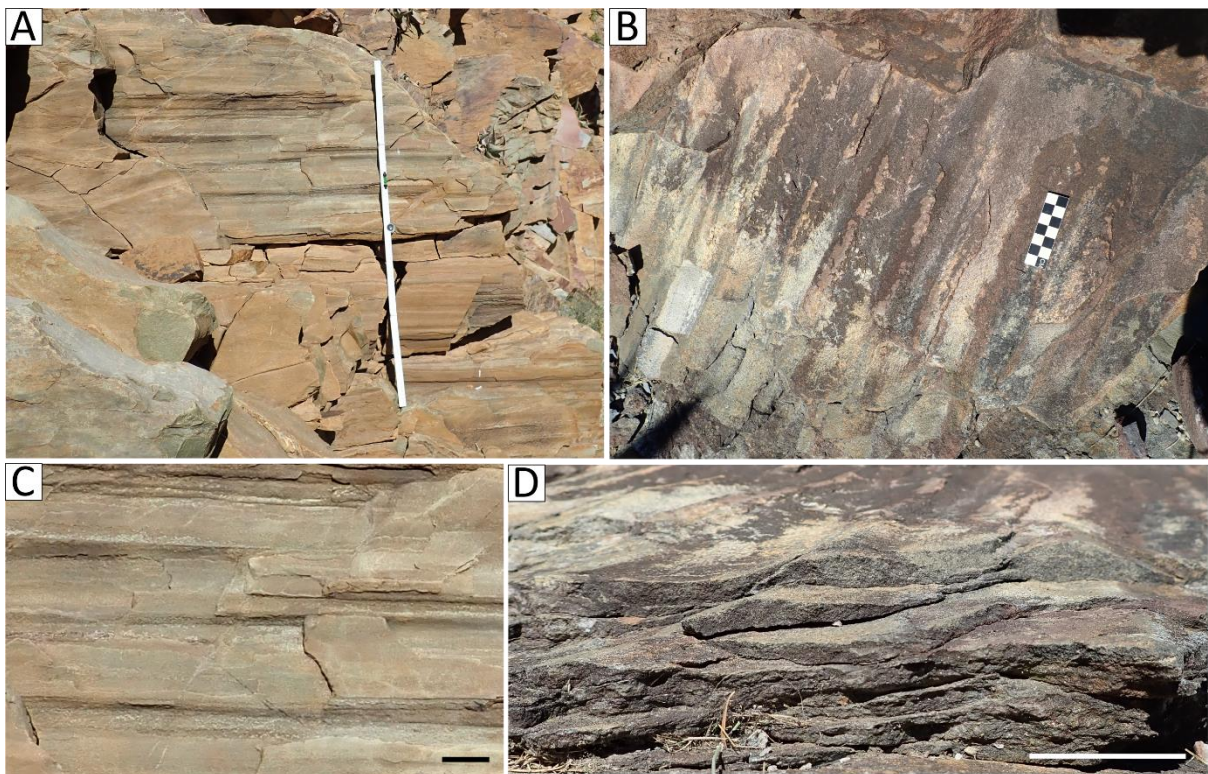


Figure 5.21: Possible ice keel scours from similar horizons in the upper Rooirand Member at Waaipoort (A, enlarged in C) and Waterloo Farm (B, D). D shows a profile view of the bedding surface in B, perpendicular to the long axis of the striae. Scale bars: A = 1 m, B = 7 cm, C = 1 cm, D = 5 cm.

Interpretations

Simple parabolic cross-stratification is the most common type of SSDS in the study area, but it commonly occurs with more complex deformations. Mountain (1964) recorded several occurrences of this type of deformation ('recumbent cross-bedding') which he considered to all occur at one horizon about one hundred feet (~30 m) below the top of the Witpoort Formation. This analysis shows that the deformation occurs at many horizons in the Witpoort Formation, including in its lower part, but are concentrated in its upper strata, particularly in the Perdepoort Member.

Soft-sediment deformation structures (SSDS) occur from liquefaction or fluidisation of unstable sediments, but the trigger cannot be diagnosed from any particular form of SSDS alone (Owen and Moretti 2011) and thus circumstantial evidence is usually required for ancient examples. This includes the sedimentological context, the areal extent of the deformation structures and their vertical relationships in succession (Seilacher 1969, 1984; Owen and Moretti 2011; Bhattacharya and Saha 2020). Following the approach of Bhattacharya and Saha (2020), determining the likely mechanism of deformation is done by a process of eliminating potential triggers. Known causes of SSDS listed by Owen and Moretti (2011) are storm waves, tsunamis, tidal shear, sediment loading, periglacial processes, meteorite impact, groundwater seepage, and seismicity. Several of these triggers, such as storm waves, tsunamis and tidal shear may be ruled out immediately for the bulk of SSDS in the Witpoort Formation because of its repetitive occurrence in cross-stratified sandstone facies that lack evidence of tidal deposition or storm waves. Similarly, SSDS related to groundwater seepage is usually caused by impermeable barriers to water escape in the

subsurface (Owen and Moretti 2011). Such impermeable layers act as a pressure valve to the movement of groundwater, leading to pressure-release deformations. A correlation between occurrence of SSDS and facies changes, such as repetitive association with overlying argillaceous material are required to invoke such a trigger (Owen and Moretti 2011; Bhattacharya and Saha 2020). Meteorite impacts may also be excluded because SSDS in the Witpoort Formation does not conform to the types of deformation expected from extra-terrestrial impacts (Alvarez *et al.* 1998) and because meteorite impacts are generally discreet events rather than repeated phenomena (Wheatley *et al.* 2016). As the described SSDS are most recognisable in cross-stratified sandstones, the depositional environment is characterised by the incremental growth of bedforms rather than by mass gravity flows or slumping (Bhattacharya and Saha 2020). Some, but not all, of the recorded deformation may be explained by loading.

This leaves two remaining triggers from Owen and Moretti's list; periglacial processes and seismic shock. Much work has been done on developing a criteria-based approach to identifying seismites in the rock record (e.g. Seilacher 1969, 1984; Owen and Moretti, 2011; Moretti and van Loon 2014; Bhattacharya and Saha 2020). According to Owen and Moretti (2011), seismites should occur 1) in large areal extents as 2) laterally extensive deformation horizons and should be 3) vertically repeated in sedimentary sequences. The SSDS should 4) be morphologically comparable to known earthquake generated liquefaction. Association of SSDS with 5) known faults and 6) zonation of intensity with respect to said faults are expected for seismites. The structures described here fulfil criteria 3 and 4 in that they occur in comparable size and form to known seismically generated SSDS (Bhattacharya and Saha 2020) and are repeated in vertical successions, separated by non-deformed beds. Criteria 1 and 2 are neither confirmed nor excluded by the present analysis. Their lateral extent is not presently well constrained; some of the deformation horizons exhibit pinch out over metres,

whereas other horizons extend for tens of metres, beyond the limits of outcrop scale observation. Lateral extensiveness of individual deformation horizons is, however, not a fundamental aspect of seismically generated SSDS (Moretti and van Loon 2014). A seismic trigger for the SSDS in the Witpoort Formation cannot be confidently inferred because there is no associated evidence for syn-sedimentary faults (aside from small scale faults at one horizon), leaving criteria 5 and 6 unfulfilled.

Based on the, albeit limited, evidence for frozen substrates, such as transported sandclasts and dropstones (5.2.10) in the study area, periglacial processes must be considered. The bedding-parallel grooves described above bear a strong resemblance to ice keel scour marks (Veseley *et al.* 2014). It has been hypothesised that glacial diamictites occur at various horizons in the Witpoort Formation (Almond *et al.* 2002), but there is no systematic study of the deposits proving a glacial origin. Recently, for the first time, soft-sediment deformation of sand was reproduced in the laboratory by liquefaction during unfreezing, producing small-scale structures such as diapirs, dykes and chevrons (Bertran *et al.* 2019). Liquefaction due to freeze and thaw therefore cannot be ruled out, however, there is presently no evidence to suggest that parabolic cross-stratification can form in this manner.

Finally, one potential trigger not considered by Owen and Moretti (2011), but which has been repeatedly invoked as a cause of parabolic cross-stratification, is drag by high-energy currents during rapid changes in flow regime (Hendry and Stauffer 1975; Røe and Hermansen 2006; Whitmore *et al.* 2015). This provides a further plausible mechanism that could explain much of the SSDS recorded here, but it would not provide a trigger for the clastic dyke in the Waterloo Farm section. This dyke is the only known structure of its kind in the Witpoort Formation, although geometrically similar clastic dykes in the Kweekvlei Formation, immediately overlying the Witpoort Formation, were illustrated by Browning and Reid (2017). Determining the mechanisms that caused these clastic dykes, whether seismic or

glacial, or some other unknown cause, may provide an explanation for much of the SSDS that occurs in the Witpoort Formation.

5.3. Sediment dispersal patterns

Palaeoflow vectors from the study area are contextualised in various ranks beginning with the broadest categories of stratigraphic unit and geological section (Fig. 5.22A and B). The flow vectors are ranked according to formation and member; the upper Weltevrede Formation and the Witpoort Formation (Rooirand Member and Perdepoort Member), but the Skitterykloof Member was excluded from this analysis due to insufficient data. The first thing to note is that there appears to be a polar shift in mean sediment dispersal direction from the upper Weltevrede Formation to the Witpoort Formation. The upper Weltevrede Formation has a polymodal distribution of vectors with a mean of 6° north. The splay of data in this unit is high with one standard error of 72° . In contrast, the Witpoort Formation has a predominating signal of southwest directed flow vectors. The Rooirand Member and Perdepoort Member both give an approximately unimodal southwest signal with mean vectors of 219° and 237° respectively and both these signals have a high consistency of one standard error less than 10° . It therefore appears that an overall reversal of sediment dispersal direction occurs from the Weltevrede to the Witpoort Formation though the signal may be affected by the dataset size, in that the upper Weltevrede Formation is represented by a much smaller number of datapoints ($n=31$).

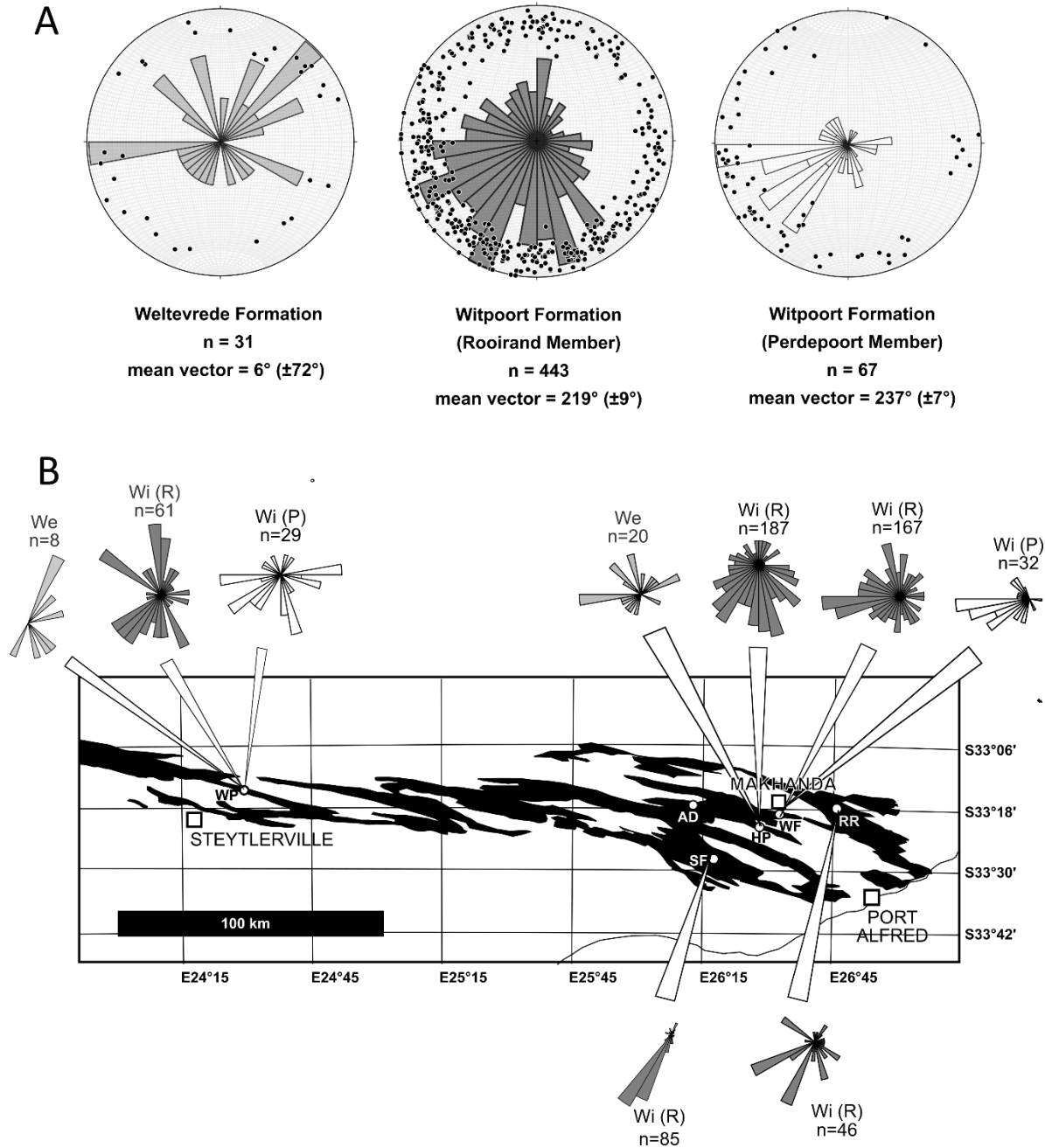


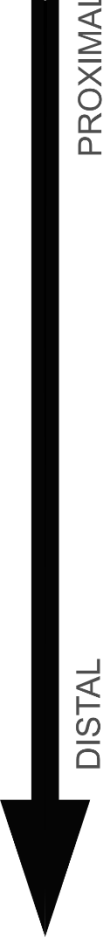
Figure 5.22: A) Palaeoflow vector roses by lithological unit. Mean vectors are rounded to the nearest degree. Error margins to mean vectors represent 1 standard error. Black dots represent individual data points. Abbreviations for geological sections are (from west to east), WP – Waaipoort; AD – Alicedale; SF – Seven Fountains; HP – Howison's Poort; WF – Waterloo Farm and Grey Dam; RR – Rabbit Ridge. Rose diagrams are shaded according to geological unit, upper Weltevrede Formation (We); Rooirand Member (Wi (R)), Perdepoort Member (Wi (P)).

To understand their variation over about 230 km along strike, the vectors are illustrated in rose diagrams by geological section (Fig. 5.22B). In this representation the signals are more

complicated. The upper Weltevrede Formation is polymodal in both the east and west of the study area, but the two sections analysed show very different trends. The strong NNE component seen at Howison's Poort is not recognised at Waaipoort, however, both sections have a strong component toward the NE quadrant, which accounts for the NE signal shown for the Weltevrede Formation in Figure 5.22A. In all Rooirand Member sections the bulk of data is directed to the SW quadrant. The Waaipoort section in the far west of the study area, however, has an additional mode towards the north, with a more bipolar signal than the Makhandia sections. Another point of interest is the difference in vector orientations between two upper Rooirand Member sections in the east of the study area, Seven Fountains and Waterloo Farm, situated 30 km apart, in a NE–SW direction, which is approximately parallel to the modal sediment transport direction for the member (Fig 5.22B). The Seven Fountains section has a strongly unimodal, SW flow direction, whereas vectors from the Waterloo Farm sections show a polymodal distribution with a predominating WSW signal. If, as hypothesised in the previous chapter, these sections represent equivalent stratigraphic horizons, this variance potentially reflects transport patterns along a proximal–distal profile. The Perdepoort Member in the east of the study area (Grey Dam section) has a unimodal WSW vector distribution, whereas at Waaipoort the vectors are polymodal, with components to WSW, SSW and ENE. To better understand these signals and the depositional environments they represent, the dataset will be analysed according to the inferred environments of deposition below.

5.4. Facies assemblages

Table 5-2: The main characteristics and palaeoflow vectors for each of the facies assemblages described in this chapter

	FA 1 Braided river sandbars (facies St, Sp, Sm, SSDS Gi, locally Ge)	Medium cross stratified beds, medium to thick structureless beds High degree of amalgamation, monotonous stacks up to 10's of m thick Laterally extensive bedsets, 10's of m long. Lack of body fossils (rare plant axes) and trace fossils Flow vectors unimodal offshore		n=140
	FA 2a Fluvio-estuarine channels and sandbars (facies St, Sp, Sc, Sl, Sm, Sh, Gi, SSDS, Sr)	Medium to very thick, cross-stratified beds with strong partings (presence of mud drapes) Rippled, silty toe sets (upwards coarsening beds) Rippled and muddy interbeds Lateral restriction Lateral facies changes Intra-unit flow vectors unimodal (offshore and offshore oblique) or bimodal (shore-parallel)		n=310
	FA 2b Estuarine sand flats (facies Sm, Sh, Su, Sc, Sp, St, Sl, Sr, SSDS, Fh)	Thin to medium tabular or wavy beds Mud drapes Heterolithic bedsets, upwards thinning, muddier upwards Abundant trace fossils (depauperate <i>Cruziana</i> ichnofacies) Flow vectors polymodal Laterally and vertically gradational to 2a and 2c		n=92
	FA 2c Estuarine basin (facies Fh, Sh, Su, Sr)	Mudrock dominated, laterally extensive lenses Ripple cross-laminated interbeds (bimodal flow vectors) Abundant plant matter, local concentrations of invertebrate and vertebrate body fossils Diminutive, poorly preserved trace fossils		n=24
	FA 2d Outer estuarine channels and sandbars (facies St, Sc, Fh, Sr)	Similar criteria to 2a, but with landwards directed cross-strata, more prominent mud drapes and muddy interbeds		n=23

5.4.1. Braided river deposits (FA 1)

Description— FA 1 comprises high energy deposits of prograding dunes and, to a lesser extent, laterally accreting dunes (Facies St and Sp beds) and sandbars formed by compound dunes (St and Sp bedsets), upper flow regime and transitional deposits (Facies Sl and Sm), and lag deposits (Facies Gi). These facies form laterally extensive, tabular units with a high degree of amalgamation and a paucity of fine-grained sediment. Intraclast conglomerates are concentrated at the base and lower portions of FA 1 units, and are usually associated with

structureless beds. Fossils are rare in FA 1 and are limited to uncommon, poorly preserved isolated plant axes, whereas ichnofossils are not recognised.

This assemblage of facies is illustrated Figure 5.23, showing its main features in sequence. This example, which is from the upper Rooirand Member at Seven Fountains represents the lower portion of a typical sedimentary cycle, such as are prevalent throughout the member.

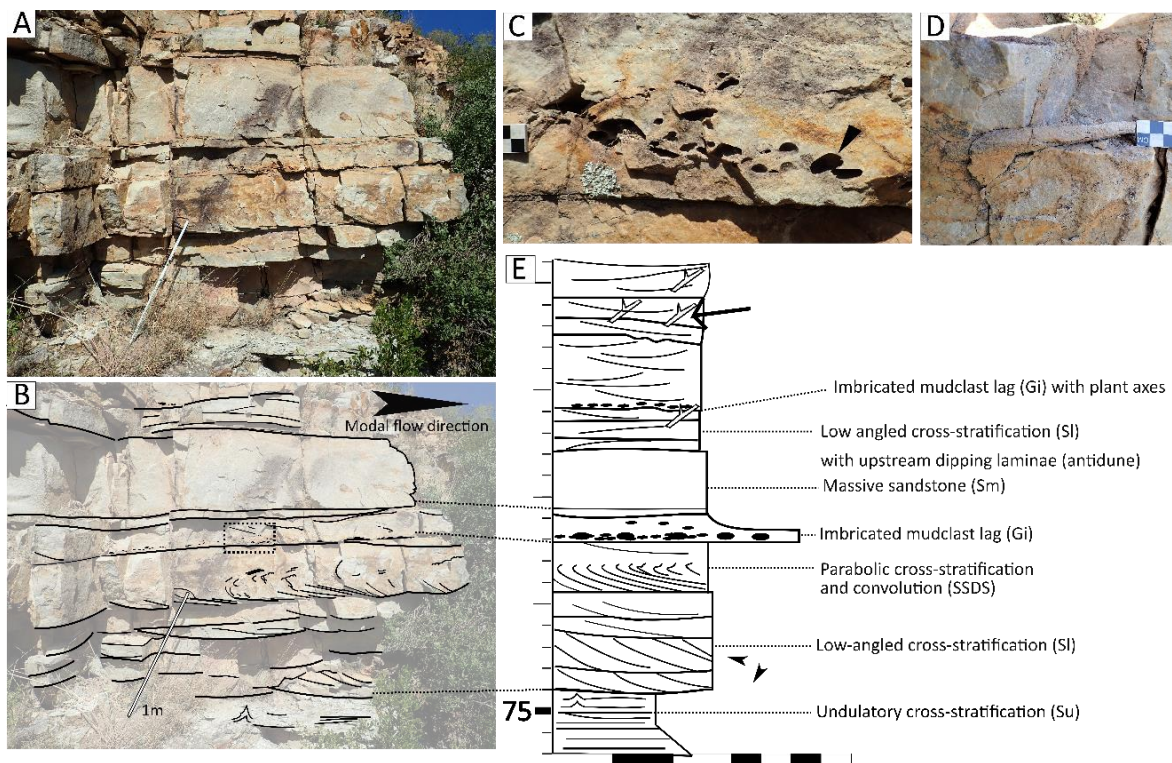


Figure 5.23: FA 1 from section SF. A) outcrop photograph with interpretive drawing (B) and stratigraphic log (E). The modal flow direction for the succession is to the right of the photograph. B) C) Mudclast stringer with imbricated clasts (indicated by arrow). The location of this photograph is shown with a stippled rectangle in B. D) Profile view of a plant axis from the horizon indicated with an arrow in E. This is probably an arborescent lycopod. E) Stratigraphic section with the facies labelled.

Low degrees of accommodation space are indicated by the amalgamated, tabular strata and the intraformational conglomerate. Normally at the base of an FA 1 succession, but also at intervals throughout such sequences, the conglomerates show that degradation of the landscape has removed and reworked prior accumulations of mud-rich strata (FAs 2b and 2c).

Amalgamated bedsets on the order of 1–2.5 m thick and tens of metres in extent represent sheetlike sandbar deposits made of medium-bedded, cross-stratified or structureless, texturally mature sandstone facies. Their bounding surfaces are approximately parallel, but may have several metre long undulations. Lamination is weakly developed, often marked by accumulations of heavy minerals along cross-strata foresets. Erosional downcutting, particularly in flow-transverse exposures, is recognised in several sections in the study area (Fig. 5.24A–C). Flow vectors for this facies association indicate predominant downstream (southwest) and lateral accretion, with a component of upstream accretion (Table 5-2).

Soft-sediment deformation structures, including parabolic cross-stratification and associated convolutions are common in FA 1 (Figs 5.23B and E, 5.25B). As discussed in 5.2.12 (above), the deformation could be related to seismicity, dewatering due to freeze and thaw, or by hypersaturated flows associated with rapid changes in flow regime (ie. floods). The last explanation is supported by experimental evidence and is highly plausible given the high-energy deposition that characterises FA 1.

FA 1 is accumulated in lithologically monotonous stacks, metres to tens of metres thick in which fine-grained sediments are limited to stringers of intraformational mudclasts. In the Perdepoort Member, FA 1 is stacked tens of metres thick (Fig. 5.25). It is divided into units that are 5–15 m thick, which extend for hundreds of metres along strike. The units are separated by laterally extensive notches in the outcrop formed by the negative weathering of slight accumulations of fines at the base. Each unit shows greater amalgamation upwards. Intraclasts are extremely rare in this member, indicating that accumulation of muds was limited in the depositional environment. FA 1 may also form accumulations tens of metres thick in the middle Rooirand Member (Fig. 5.26), although intraclast conglomerates and stringers are almost always evident in these successions (ie. WP 1 (134–152 m; Appendix 1C) and HP 3–4; Appendix 1H). More commonly in the Rooirand Member, FA 1 forms thinner

units (metres thick) overlying intraclast conglomerate, which pass up almost imperceptibly into FA 2, with decreasing amalgamation, lateral restriction and the appearance of mud-drapes (see 5.4.2 below).



Figure 5.24: Erosional incision (indicated by stippled lines) associated with FA 1: A) A concentrically filled scour from the upper Rooirand Member at Seven Fountains (ruler is 1 m long). B) Erosive channel cutting with approximately 1 m of erosional relief, filled with amalgamated trough cross-stratified quartz arenite, middle Rooirand member at Howison's Poort

(HP 3). C) Deep channel erosion with massive infill, upper Rooirand Member at Waaipoort (WP 1). Depth of incision was not measured due to inaccessibility of the outcrop.

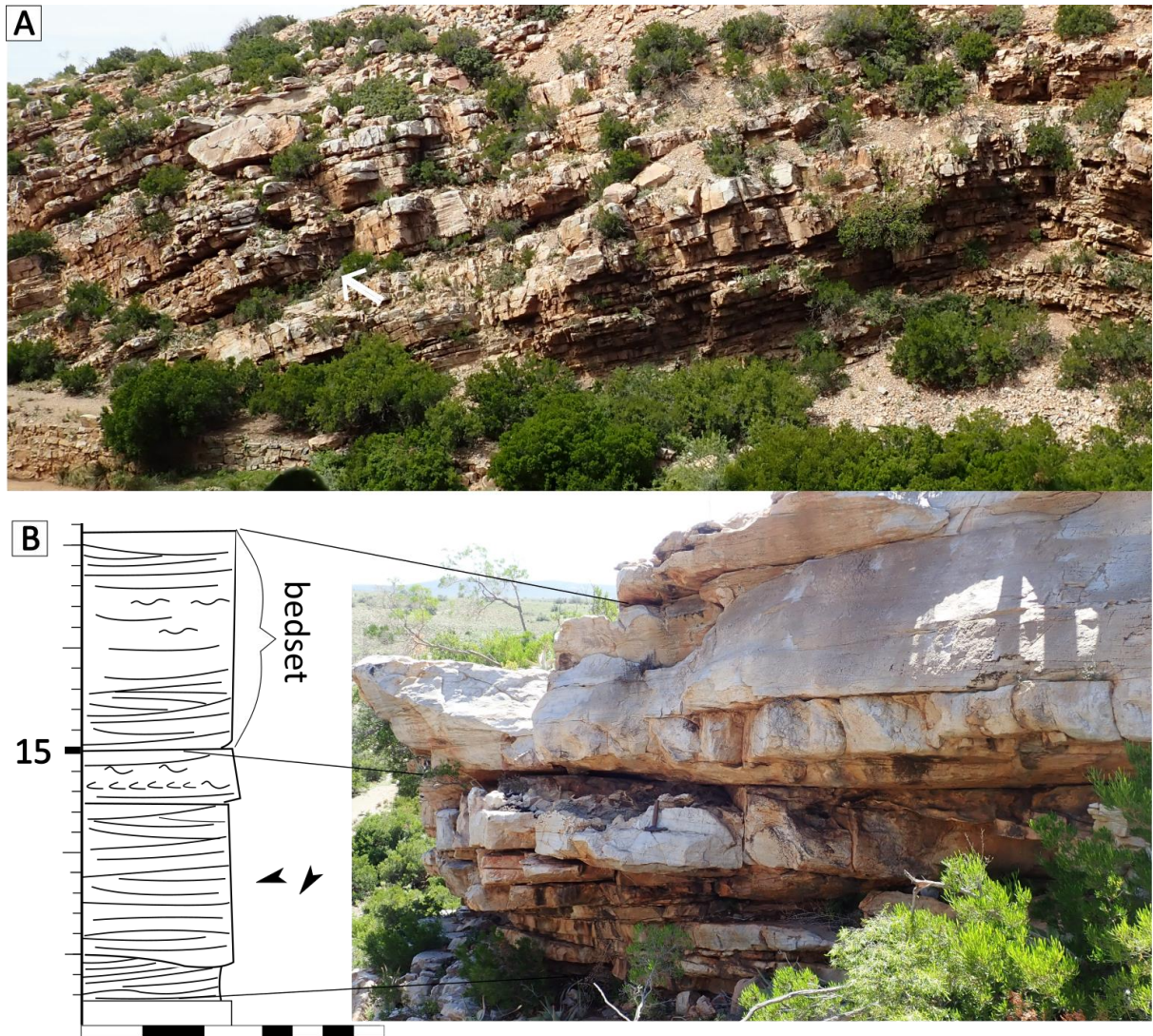


Figure 5.25: FA 1 in the Perdepoort Member at Waaipoort (WP 2). A) Multiple stacks of tabular FA 1 units (5-15 m thick) separated by slight notches with negatively weathered finer grained sandstone at the base of each unit. The position of the outcrop shown in B is indicated with an arrow. B) The lower 5 m of a unit, with a notch at the base, and strata increasingly amalgamated upwards. (Geological hammer for scale)

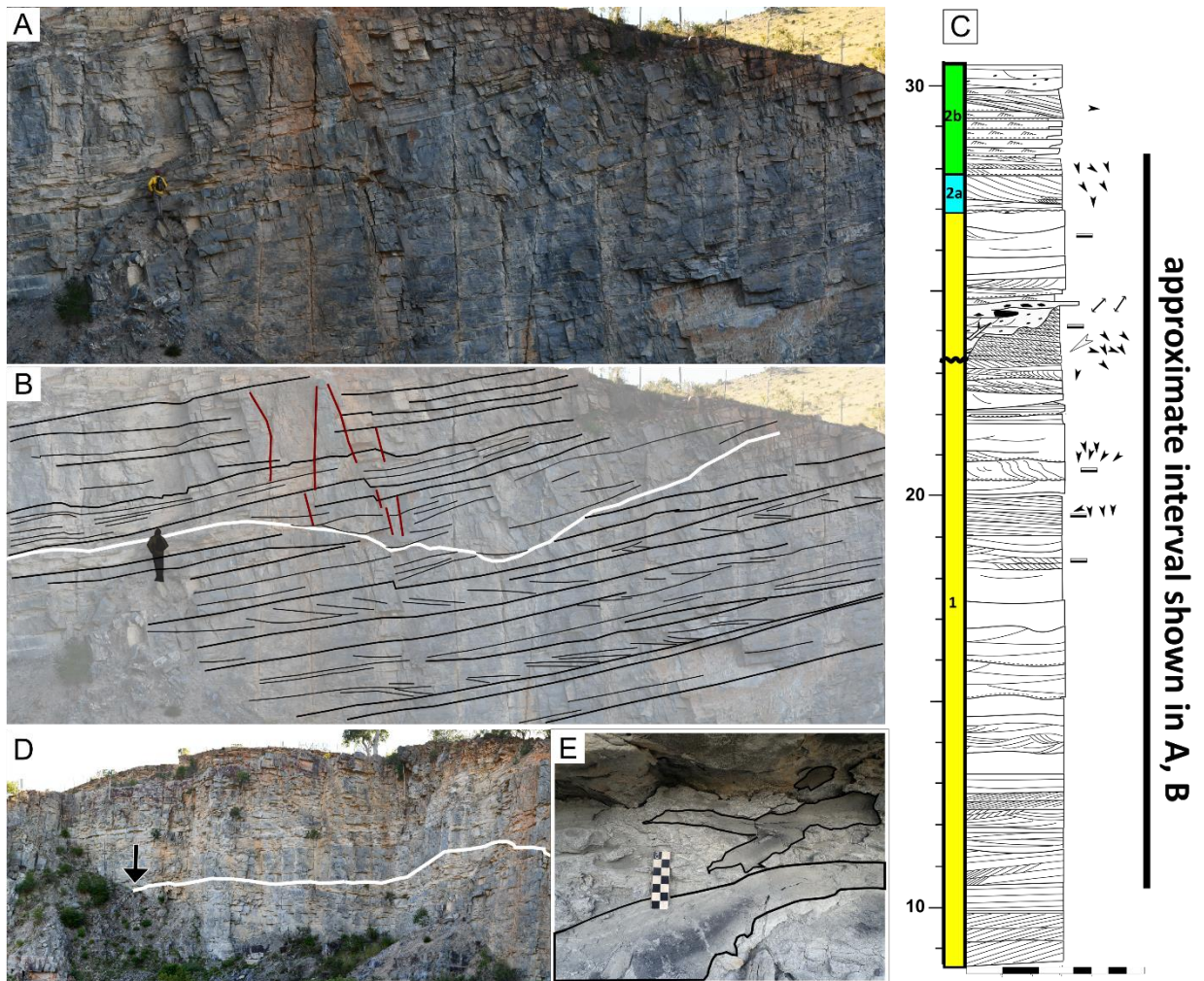


Figure 5.26: An FA 1 succession stacked tens of metres thick in the middle Rooirand Member at Howison's Poort (HP 4). A) Outcrop photograph with interpretive drawing (B), person for scale. Note the tabular, laterally extensive stratal geometry (black lines). The prevailing current direction is to the right of the photograph. A deeply incised, laterally extensive erosion surface (thick white line) separates FA 1 packages. Bedding perpendicular fractures are indicated by red lines. C) A stratigraphic log, measured approximately 50 m northwards along strike from photograph A (scale increments are in metres, Facies Assemblages are indicated at the left of the column). The erosion surface occurs at 23–24 m in the logged section, where it is locally overlain by intraclast conglomerate (black arrow in D). E) Boulder-sized mudclasts resting directly above the erosion surface (scale increments are 1 cm).

Interpretations– Thick, lithologically uniform accumulations of tabular sandstones are typical for braided river systems (Miall and Arush 2011), of which many ancient examples bear a resemblance to this facies association. Some relevant examples include the Late Jurassic Bencliff Grit (Allen and Underhill 1989), the Lower Devonian Brownstones of the Welsh borderlands (Allen 1983) and the Upper Devonian Toe Head Formation (Cotter and Graham 1991), the Ordovician Peninsula Formation in South Africa (Turner *et al.* 2011), the Devonian Battery Point Formation (Cant and Walker 1976). Braided rivers develop in fluvial systems with high discharge variability, such as occur in strongly seasonal and arid regions (Fielding 2006; Miall 2010; Colombero *et al.* 2013; Fielding *et al.* 2018; Fielding and Best 2019).

An open-marine genesis may be excluded on the basis that it is inconsistent with the facies and stacking patterns, unimodal current roses (Table 5-2), and the lack of marine body and ichnofossils characterising FA 1. Clastic shoreline facies models predict that the thickness of shoreface deposits increases with increasing wave and storm energy, and should reflect typical upwards-shallowing parasequences not exceeding 10 m thick in the rock record (Clifton 2006, Plint 2010). Forced regression may cause the amalgamation of shoreface sandbodies, increasing their thickness (McCrory and Walker 1986; Walker and Bergman 1993, Hampson and Storms 2003; Dalrymple and Choi 2007; Plint 2010), however, nowhere in the study area is the typical facies progression of a high-energy shoreface recognised. Open-marine shorefaces are controlled by waves, storms and alongshore currents. Below fair-weather wave base, periodic storm deposits are preferentially preserved, producing hummocky and swaley cross-stratification (Clifton 2006). The intermittent nature of these event beds enables the colonisation of the substrate by marine benthos during quiescent intervals (Buatois and Mángano 2011). Upper shoreface deposits should have evidence of alongshore currents as well as gently dipping seaward and landward planar lamination on the

foreshore, and typically host suspension-feeding burrows (Plint 2010; MacEachern *et al.* 2010). None of these characteristics are recognised in FA 1.

Perhaps the clearest idea of what the various facies comprising FA 1 signify may be derived from neosedimentological studies on braided river systems. In ancient outcrops the great lateral extent of individual units makes separation of the various geomorphological elements, such as, for example, the identification of individual channel bodies, challenging (Miall 1985). In modern braided rivers the main geomorphological elements are sandbars, dunes, channels, flats, overbank muds and vegetated islands (Collinson 1970; Cant and Walker 1978; Allen 1983; Miall 1977, 2010). Deposits with the highest preservation potential are the downstream parts of channel bars formed during flooding (Bridge 1993). Linguoid bars are predominating architectural elements in certain modern braided river deposits such as the Tana River (Collinson 1970). They form downstream and laterally accreting bodies 10's to 100's of m long and wide, with a convex downstream margin and tend to form trough cross-stratified sets (Collinson 1970). Braided river systems with higher discharges favour the formation of sand flats over linguoid bars (Cant and Walker 1978; Allen 1983). Low angled cross-stratification (Sl) with concave and convex up foresets resemble the sand flats facies of Allen (1983), the laminated sand sheets architectural element of Miall (1985) and the humpback dunes of Bridge (2003) and Fielding (2006). In all of these studies, high energy, shallow water unidirectional flows (upper flow regime, or transitional lower to upper flow regime) were invoked. Bridge (2003) and Fielding (2006) suggested modifying Miall's (1985) laminated sand sheets to an upper flow regime architectural element, considered to be indicative of strongly seasonal, flashy discharge. This is because rapidly changing flows do not allow beds to equilibrate, which enables the preservation of peak flood event beds. In such deposits, parting lineations should be common (Fielding 2006), however, their absence can be explained by the etching and removal of surface textures by wind deflation of exposed

sandbars (Cant and Walker 1978). Planar cross-stratification forms in straight-crested dunes, and in the context of braided river systems, this has been attributed to cross-channel sandbars (Miall 1977; Cant and Walker 1978; Allen 1983). The upward thinning pattern regularly seen in Sp bedsets is typical for this architectural element, which is accounted for by shallowing of the bar top as it aggrades (Cant and Walker 1978).

An important characteristic of FA 1 is that it may form great thicknesses with no apparent record of overbank fine-grained strata preserved. Devonian and younger braided river systems regularly preserve overbank deposits, whereas older, pre-vegetation systems show a paucity of overbank muds (McMahon and Davies 2018). This has been explained by bypassing of fines due to aeolian winnowing, because baffling by vegetation was non-existent (Dalrymple *et al.* 1985; Davies and Gibling 2010). Apart from plant axes associated with conglomeratic horizons (reworked FA 2), FA 1 provides no evidence for vegetated landscapes, and therefore, alluvial strata in the Witpoort Formation may represent an analogous scenario; a strongly seasonal discharge regime characterised by periodic flooding and harsh subpolar weather may not have favoured luxuriant vegetation. On the other hand, FA 1 is highly amalgamated and, if present at the time of deposition, vegetation, roots and soils will have had a low preservation potential in these facies. The non-recognition of typical aeolian deposits in these strata does not exclude the possible importance of aeolian processes as subaerial deposits have a low preservation potential in the rock record (Davies and Gibling 2010). Adhesion ripples found in the Seven Fountains section provide the first evidence of deposition by wind from the Witpoort Formation, and it is not excluded that some of the strata could be aeolian.

5.4.2. *Estuarine deposits (FA 2)*

An estuary is classically defined as a river mouth or inlet of the sea in which sea water is diluted by fresh water from land drainage (Pritchard 1976; Day *et al.* 2012). For ancient

deposits however, the distribution of lithofacies does not entirely reflect salinity levels and hence, from a geological perspective, Dalrymple *et al.* (1992, p. 1132) defined an estuary as “a drowned valley system... extending from the landward limit of tidal facies at its head to the seaward limit of coastal facies at its mouth.” Subsequently, Dalrymple (2006) expanded this definition beyond drowned valleys to include transgressed portions of delta plains.

Facies assemblage 2 occurs in vertical succession with facies assemblage 1, which it typically overlies conformably in the study area. The transition from FA 1 to FA 2 is, in many cases, cryptic and gradational (Figs), and is indicated by the presence of increasingly lenticular sandstone bodies, abundant clay, ripple cross-lamination and the preservation of trace and body fossils indicative of brackish water environments. Based on the lithofacies distributions, flow vector orientations and trace and body fossils, four estuarine sub-environments are recognised in the study area; fluvio-estuarine channels and bars (FA 2a), sand flats (FA 2b), estuary basin muds (FA 2c) and outer estuary tidal bars (FA 2d). Their main characteristics are summarised in Table 5-2.

These deposits occur in the overall upward-fining (muddier-upward) cycles that characterise the Rooirand Member, and the Skitterykloof Member, but are not apparent in the Perdepoot Member.

FA 2a– Fluvio-estuarine sandbars and channels

Description– This assemblage comprises planar and trough cross-stratified sandstone beds (St, Sl and Sp), commonly having mud-draped foresets and accumulations of fines in the toe-sets. In contrast with FA 1, these beds show a decrease in lateral extent and amalgamation. Cross-stratified beds are, in places, very lenticular (metres in lateral extent), and are medium to thick-bedded (occasionally very thick) with clearly defined foresets that have a strong parting. Lamination can be highly irregular in thickness, and is often developed in inclined sigma or wedge-shaped laminasets separated by slight reactivation surfaces (Figs 5.4B, 5.6E,

5.15 G and F, 5.26D, 5.27E). Ichnofossils are uncommon in this facies assemblage. They include vertical burrows of probable suspension feeders (*Altichnus*) shallow resting traces (*Selenichnites*) and plug-shaped burrows. Mud-clast stringers lining beds and foresets are common, in places containing abundant fossil plant axes. Flow vectors from this FA (n=310) are unimodal to the SW, similar to those from FA 1, indicating that the predominant transport direction is basinwards.

A good example of the transition from FA 1 to FA 2 is exposed at Howison's Poort (HP 3), where a package of tabular, medium beds is overlain by a ~2 m thick wedge-shaped unit which rapidly pinches out in the direction of cross-strata progradation (Fig. 5.27). The unit contains thick, sigma-shaped St beds with multiple internal surfaces of reactivation, irregular heavy mineral laminated and clay draped foresets with silt accumulations in the toe sets. The unit is sharply overlain by carbonaceous shale (FA 2c), containing plant fossils that were studied by Plumstead (1967).

At Waterloo Farm, this facies assemblage forms significantly thicker units, which appears to predominate in this 80 m thick upper Rooirand Member interval. A 6 m thick succession at the base of section WF 1a exposes an array of variations with this facies assemblage (Fig. 5.28). It comprises a lower unit of thick wedge-shaped sandbodies that rapidly thin and grade down-dip to ripple cross-laminated toe sets containing trace fossils (C1 in Fig. 5.28B). The lower wedge-shaped unit is overlain by a composite set of channelised, laterally accreting trough cross-strata, which has a lag deposit at the base containing arborescent lycopod axes (C2 in Fig. 5.28B). This sequence is capped by a parallel-laminated siltstone lens with comminuted plant fragments, which is underlain by a thin lag deposit of platy mudchips (FA

2c). Flow vectors in this succession appear to indicate strong progradation followed by lateral accretion and then retrogradation.

Significant variation in the expression of this facies assemblage occurs in the study area. FA 2a is present in all studied sections, often in an intermediate position between the stacks of FA1 arenite and the more heterolithic and muddy estuarine deposits (FA's 2b and 2c). Well-developed examples are observed in section at Seven Fountains (Fig. 5.29C), in the interval immediately underlying the Waterloo Farm upper shale (WF 2, Figs 5.29A, 5.15D and E) and at multiple horizons in the Rooirand Member at Waaipoort (Fig. 5.29B). The upper Weltevrede Formation at Howison's Poort has an example of this facies assemblage that it would be appropriate to describe and illustrate in this section.

These successions are lithologically complex and require detailed investigation with high-resolution flow vector data to understand their nuances, as is evident from the Waterloo Farm example illustrated in Figure 5.27.

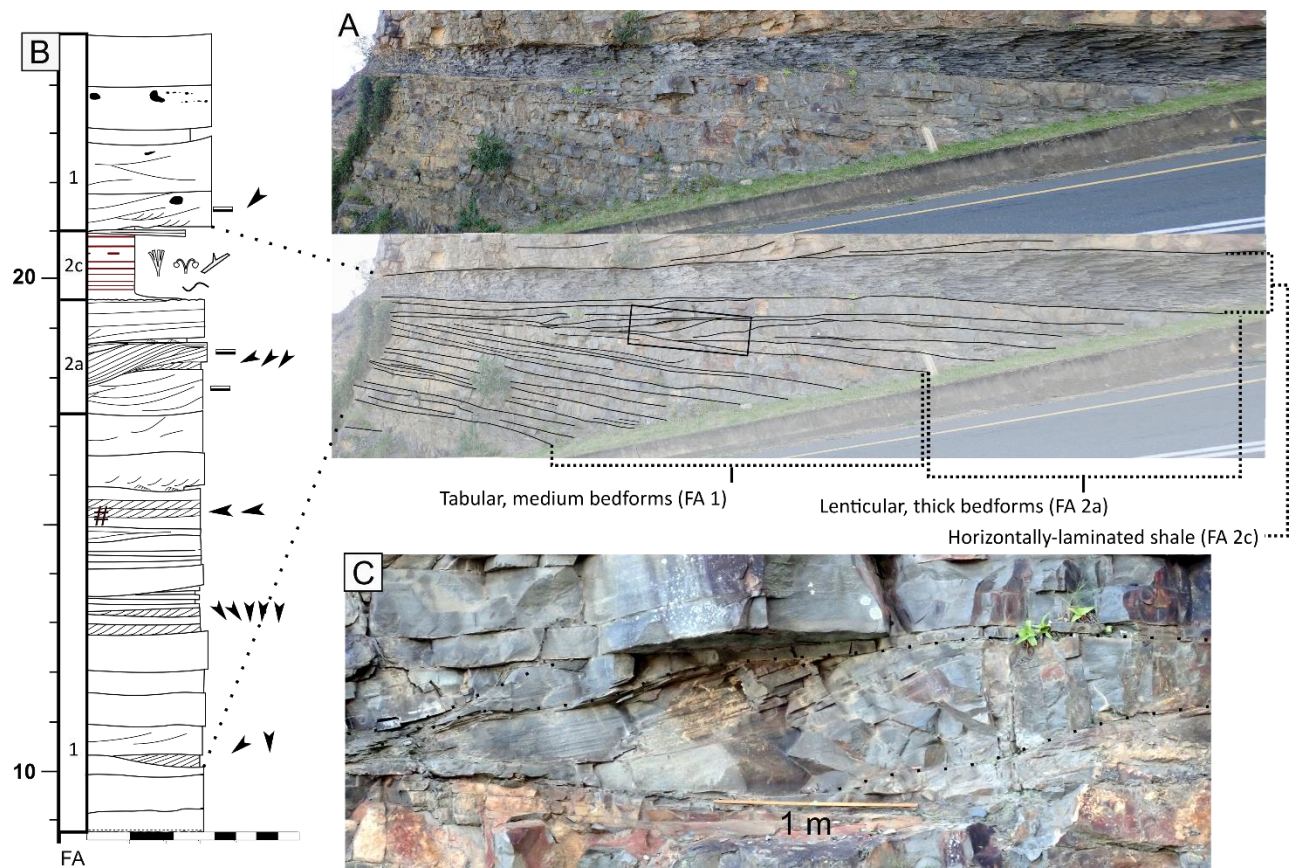


Figure 5.27: FAs 1 and 2 from the lower Witpoort Formation at Howison's Poort (HP 3). A) Outcrop photograph and line drawing showing the bed contacts. B) Stratigraphic log showing the sedimentary facies, flow vectors (black arrows) and facies assemblages. Symbols are as given in Appendix 1B. C) Enlargement of the bed shown with a rectangle in A. Reactivation surfaces within this bed are marked by fine stipples.

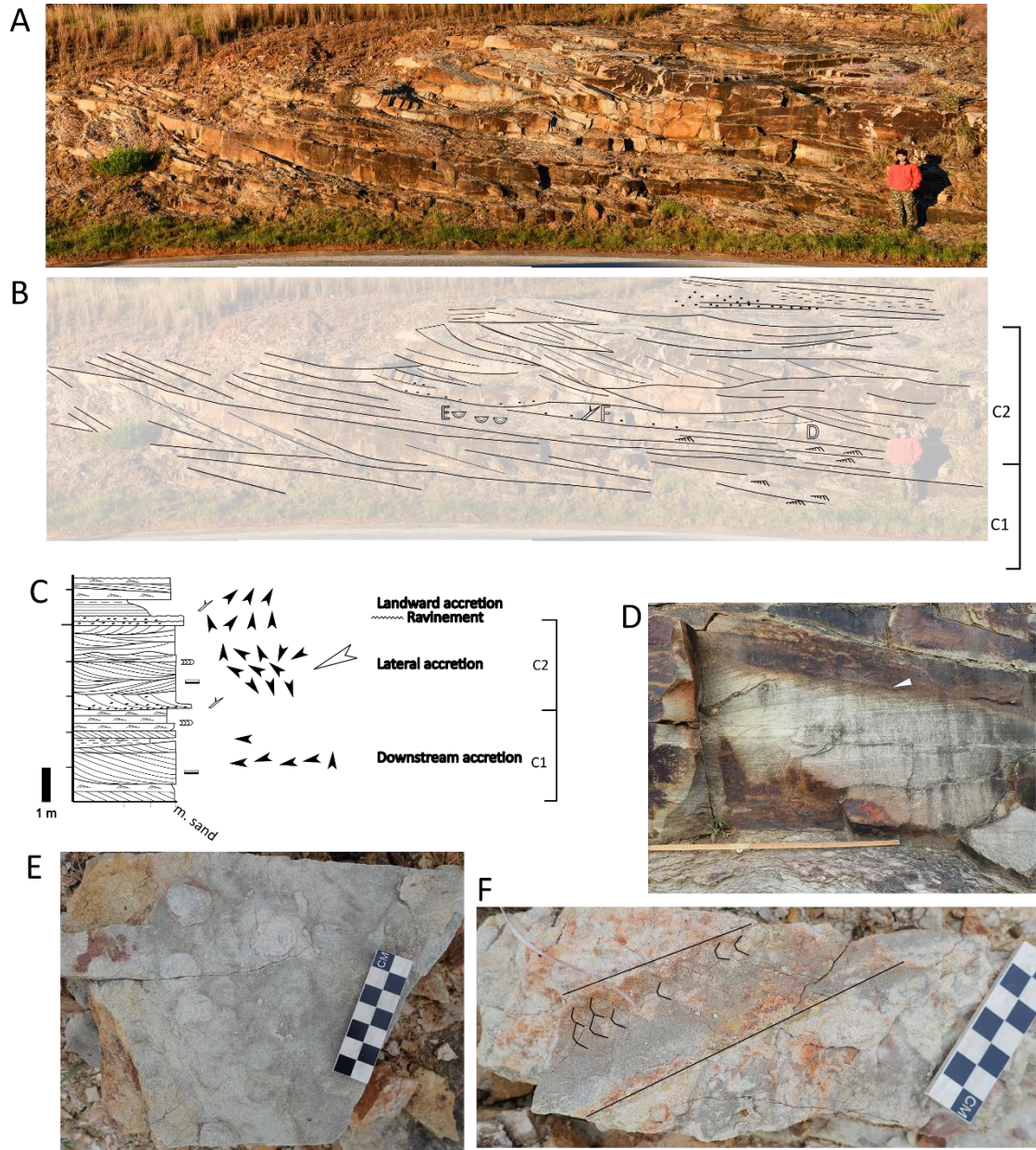


Figure 5.28: Fluvio-estuarine channels (FA2a) from Waterloo Farm (WF 1a). A) Outcrop photograph. B) Line drawing of A. Labels refer to the stratigraphic log of the outcrop (C) and photographic enlargements (D–F). The succession shows a transition from prograding wedge-shaped sand bars (C1) to lenticular, laterally migrating channel forms (C2), erosionally overlain by basinal muds (FA 2c). Enlargements from this outcrop include, D) a trough cross-stratified channel sandbar (reactivation surface arrowed). E) Plug shaped burrows at the bottom of a thin sandstone bed in a C1 toe-set. F) Plant axis (*Leptophloeum rhombicum*) with line drawing overlay showing the characteristic rhombic abscission scars.

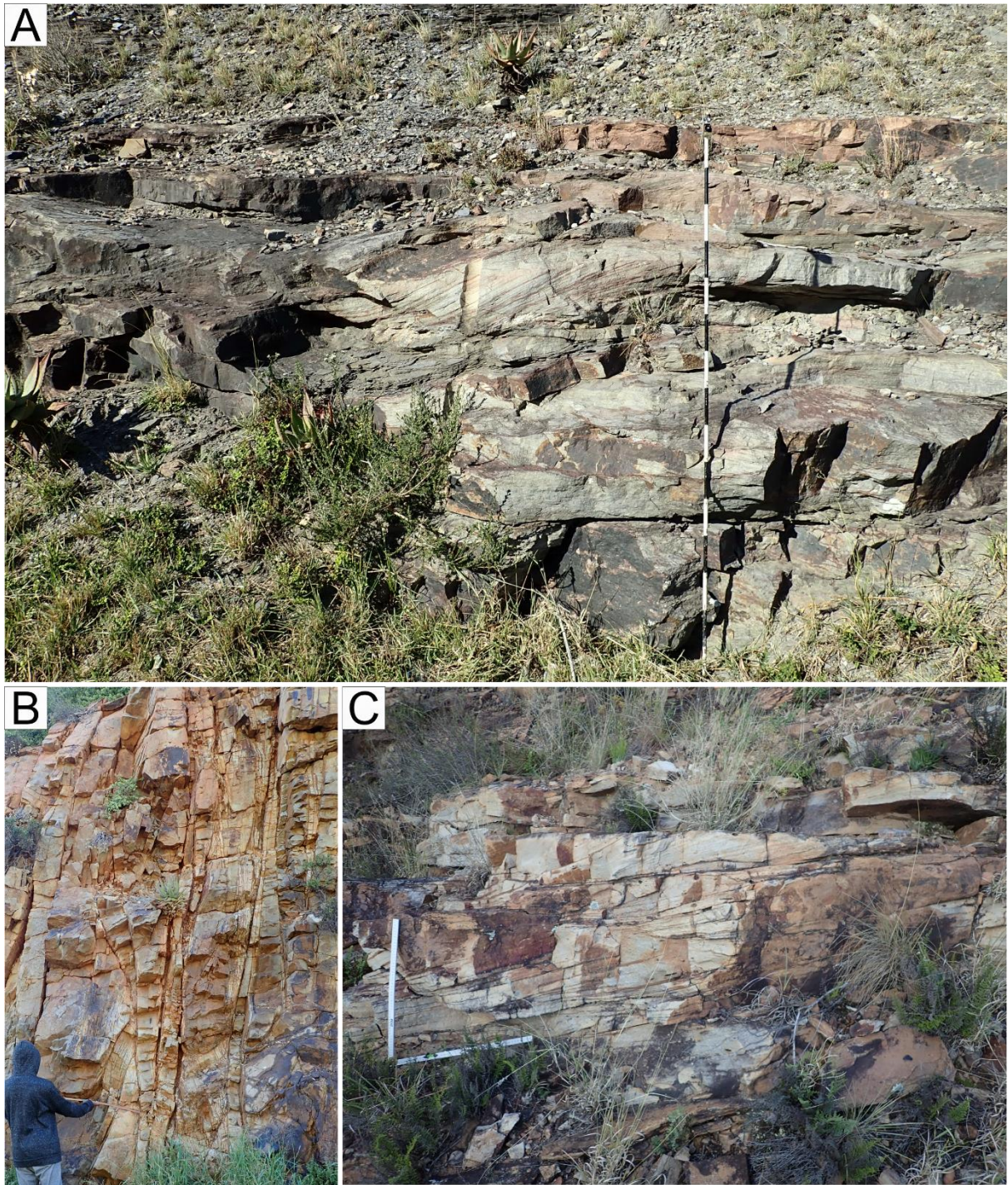


Figure 5.29: Facies assemblage 2a, interpreted as deposits of fluvio-estuarine channels. A) Immediately underlying the Waterloo Farm upper shale (WF 2, jacob staff is 1.5 m). B) Rooirand Member at Waaipoort. C) Upper Rooirand Member at Seven Fountains (ruler is 50 cm²).

Interpretations– The lateral restriction and larger scales of cross-stratification found in FA 2a indicate deeper water, more channelised flows compared with FA 1. As discussed above (5.2.3), dune height in subaqueous flows is largely controlled by the water depth (Best and Fielding 2019), but their preservation could also indicate higher rates of accommodation space generation for FA 2a. Similar palaeocurrent roses for both facies assemblages and their conformability in succession suggest that the two assemblages occurred along a continuum of laterally adjacent environments with equivalent, basinwards flow patterns. The presence of muddy partings (including double drapes), reactivation surfaces and irregular foreset thicknesses are features that can arise by tidal processes (Visser 1980; Boersma and Terwindt 1981; Van den Berg *et al.* 2007; Dalrymple *et al.* 2015). Tidal pushback influences the morphology of fluvial dunes (Zaitlin *et al.* 1994; Van den Berg *et al.* 2007), favouring planar-tabular cross-stratification over trough cross-stratification (Dalrymple and Choi 2007) and increased dune heights (Shchepetkina *et al.* 2019).

A stratigraphic position in a conformable sequence overlying fluvial deposits and underlying estuarine muds, is consistent with deposition in bayhead deltas, forming the interface between alluvial and estuarine environments (Dalrymple *et al.* 1992; Allen and Posamentier 1993; Plint and Wadsworth 2003; Aschoff *et al.* 2018). In this depositional setting, intermittent mud drapes can be explained by upstream shifts of the tidal limit during the largest tidal events (i.e. spring tides) (Dalrymple *et al.* 2015). Bayhead deltas typically produce basinward-prograding clinoforms at scales of metres in thickness and hundreds of metres to kilometres in length (Aschoff *et al.* 2018). Deposition within the fluvially controlled and tidally modulated inner estuarine environments explains the characteristics attributed to this facies assemblage. Depending on the tidal range, coastal gradient and scale of fluvial discharge, the tidal limit may extend for tens to hundreds of kilometres inland in

modern estuaries (Van den Berg *et al.* 2007; Dalrymple and Choi 2007) and it is therefore possible that FA 2a accumulated at comparable distances inland from the open ocean.

Ancient fluvio-estuarine successions in deposits of the Moodies Group (Archean) (Zametzer *et al.* 2023) and the Cretaceous McMurray Formation illustrate similar trends (Dalrymple *et al.* 2015; see also Jablonski and Dalrymple 2016, Martinius *et al.* 2015).

FA 2b– Heterolithic estuarine sand flats

This is a heterolithic association of facies representing the interfingering of the sand-dominated and mud-dominated depositional environments. It comprises very thin to medium, tabular or wavy sandstone beds of parallel laminated (Sh and Su), ripple cross-laminated (Sr), cross-stratified (St and Sp) or structureless sandstone, interbedded with argillaceous facies, ranging from a thin drape of mud to horizontally laminated shale (Fh) to flaser and lenticular ‘wavy’ bedding (Sr (a and b)) and rhythmic lamination (Sc). The sandstone beds are sheetlike, usually sharply based, and are superior in thickness to the argillite or occur as interbeds within mud-dominated successions (FA 2c) (Figs 5.31 and 5.32; also Fig. 3.3). They are typically developed in thinning upwards, muddier-upward units, but also thicken upwards in places. Stratigraphically this association of facies usually occupies an intermediate position between underlying channel deposits (FA 1 and FA 2a) and estuary basin muds (FA 2c). FA 2b is gradational to both FA 2a and 2c.

Flow vectors for FA 2b are polymodal. The sand-rich strata show a strong signal of offshore (southwest) and alongshore (northwest–southeast) sediment dispersal, whereas reversals in the flow pattern are in many instances recorded in ripple cross-laminated interbeds in the muddier parts of successions which often show evidence of waves (facies Sr (b), see 5.2.9)

This association of facies produces the bulk of the trace fossils recognised in the study area. Trace fossil distributions are very sporadic, and where occurring they exhibit low

diversity (in most cases 1–3 ichnotaxa) and low bioturbation intensities (BI 0-1). Bioturbation is largely bedding-parallel, with combined locomotion and deposit feeding as the predominant strategy. Vertically orientated traces are uncommon and are limited to restricted, often monospecific horizons. These biological characteristics are consistent with fluvio-tidal transition zones (Dalrymple and Choi 2007; Shchepetkina *et al.* 2019). The ichnotaxonomic composition and palaeo-environmental significance of these traces will be analysed in greater detail in the next chapter.

Interpretations– The sheetlike geometry of the strata indicate deposition in flats, which within the estuarine realm may occur as a result of marine washover (Reddering 1983), tidal currents (Boersma and Terwindt 1981, Dalrymple and Choi 2007), and intermittent fluvial discharge (Van den Berg *et al.* 2007; Dalrymple *et al.* 2015). Distinguishing between these processes relies on stacking patterns, palaeocurrent data and ichnology. Occupying a stratigraphic position within the transition from fluvio-estuarine channels to estuary funnel muds places certain examples of this facies association within the fluvial-tidal transition zone. The ichnological signals (Chapter 6) and flow vectors are consistent with this interpretation. The possible influence of marine washover during storms is, however, not excluded for some examples. For example, thin beds of laminated sandstone from the Rabbit Ridge upper section contain marine trace fossils, particularly *Nereites* and *Lingulichnus* indicating salinities elevated above that expected in the fresh to brackish inner estuary. This could also be the result of the upstream shift of the salinity limit during periods of low fluvial discharge enabling colonisation of fluvial sands by marine organisms (Dalrymple *et al.* 2015). An association with facies Su (5.2.7) in some outcrops provides further evidence for storm wave deposition as an influence on this facies assemblage.

Deposition of mud in the fluvial-tidal transition zone is controlled by the upstream-downstream shifting of the turbidity maximum, regulated by the interplays of fluctuations in

river discharge with systematic variations in the amplitude of the tidal cycle. These variables contribute to ‘estuarine circulation’ processes that generate upstream transport of muds via a bottom-driven salt wedge. Thus, a complex interplay of fluvial sand, tidally modulated channels and bars, and estuarine muds are generated (Van den Berg *et al.* 2007). The hypothetical interactions between fluvial and tidal processes in the fluvial-tidal transition zone are schematically illustrated in Figure 5.30 (after Dalrymple *et al.* 2015), explaining the relative importance of tidal signals over the course of a river flood event with increasing distance from the coast. In a strongly seasonal climatic regime as predicted for the Witpoort Formation, cyclical alternations between river floods and interflood periods could represent annual events (Fig. 5.31). Systematic variations in the amplitude of the tidal cycle may also explain some of the rhythmicity, producing regular alternations of sandstone and mudstone (Kvale *et al.* 1995) (Fig. 5.32). Counterintuitively, the largest tidal events (ie. spring tides) can be responsible for intermittent mud deposition due to the development of fluid muds (Mackay and Dalrymple 2011), the inverse pattern to that expected for tidal flat deposits (Choi and Park 2000).

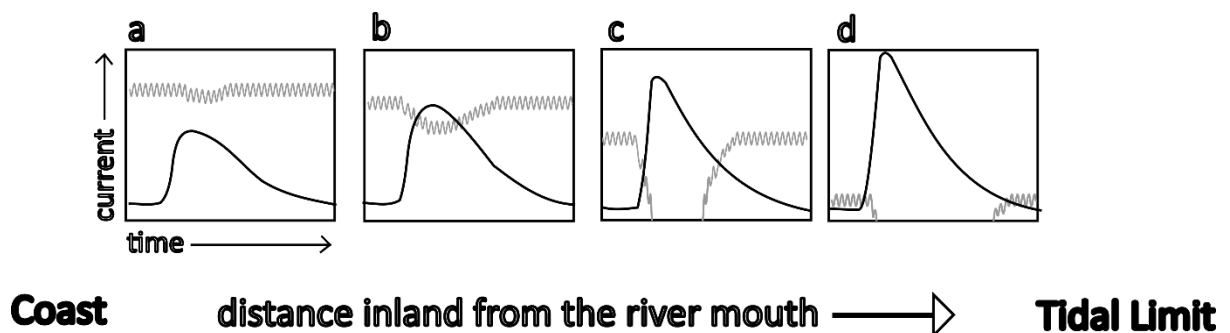


Figure 5.30: The relative importance of fluvial and tidal processes in the transition zone during a river flood event, based on Dalrymple *et al.* (2015). The diagrams (a–d) represent increasing distance from the river mouth in a macrotidal estuary between the coast and the tidal limit. The horizontal axes represent an unspecified amount of time and the vertical axes represent the relative importance of fluvial (black lines) and tidal (grey lines) currents. A location proximal to the coast is shown in (a), where tidal currents predominate. Note that the tidal currents decrease slightly during the peak of river flooding, but this is not enough to produce a fluvial sedimentary signal. By contrast, (d) indicates a location close to the tidal limit. Here the tidal signal is lost completely

during a river flood event, and only presents a weak signal during the interflood period when the tidal limit is displaced further landward. Diagrams (b) and (c) indicate locations intermediate between (a) and (d).

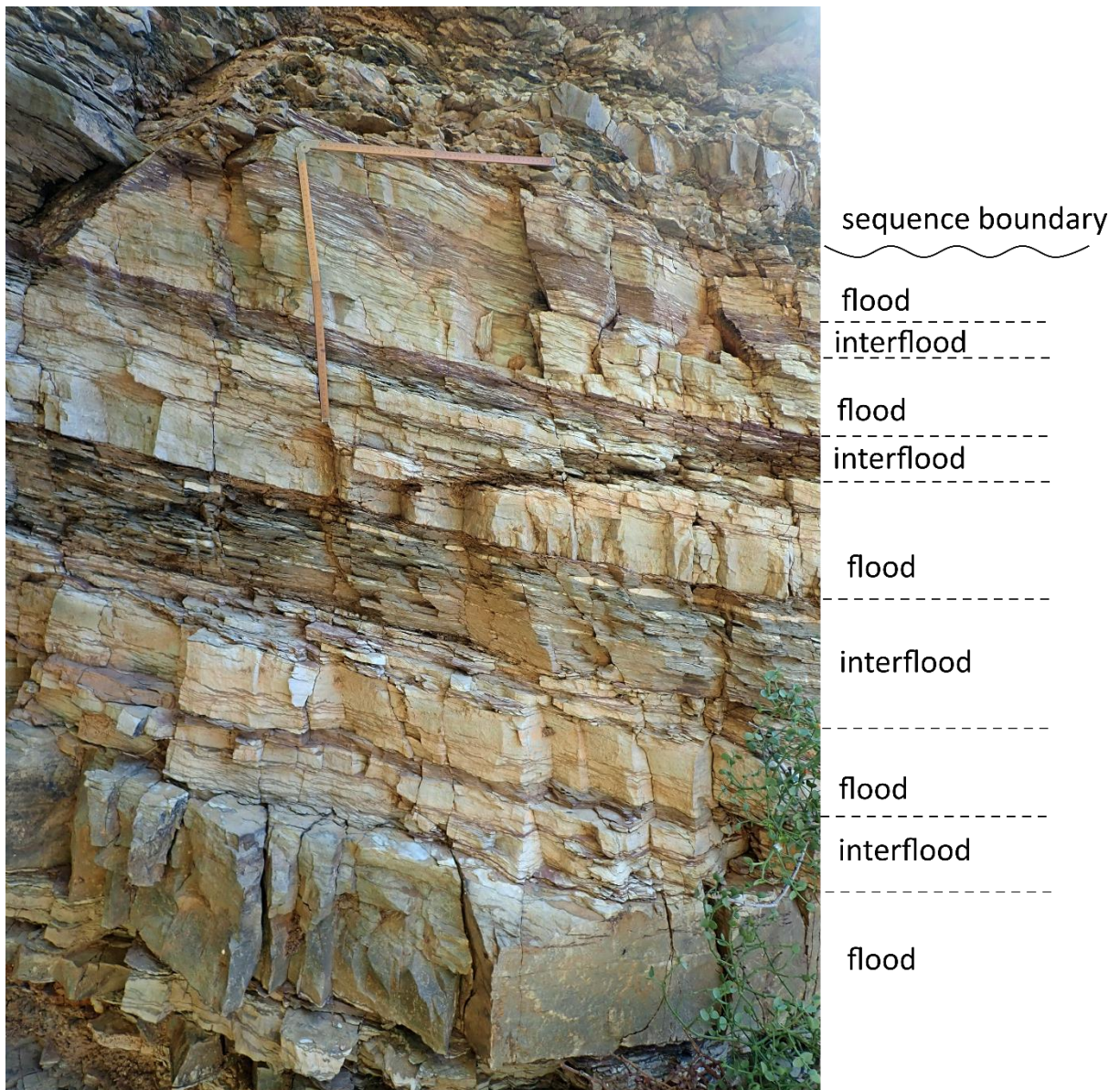


Figure 5.31: a stratigraphic sequence interpreted as deposits of the fluvial-tidal transition zone (scale is 50 cm x 50 cm). The unit contains medium beds of sharp based structureless to ripple cross-laminated sandstone (flood deposits) grading upwards to heterolithic mud-rich beds (interflood deposits). Sandstone beds contain increasing mud-drapes upwards, including paired drapes. They grade upwards into interflood deposits containing southward (offshore) prograding climbing ripples and rhythmic very thin laminae of fine sand and silty mud. Although the deposit contains no landward flow vectors, it is hypothesised that the mud-rich strata were deposited when estuarine circulation prevailed during interflood periods due to the landward shift of the tidal limit. This deposit occurs at the top of a 4 m thick upward-fining sequence, and is erosively

truncated by an intraformational conglomerate deposit containing pebble to boulder-sized shale clasts.



Figure 5.32: Thin, sharp-based sandstone beds (3–7 cm thick) separated by mud-drapes. This bedset occurs within a 6 m thick mud-dominated horizon about 90 m above the base of the Witpoort Formation at Waaipoort. Regular mud drapes (including double drapes) potentially represent the largest tidal events in a tidal cycle causing the development of fluid muds during the course of a river flood event.

FA 2c– Muddy estuarine basin ('estuary funnel')

Description and interpretations– Mud-dominated strata in the study area include very thin to very thick (up to 12 m thick) beds of horizontally laminated shale and mudstone, as well as heterolithic beds with millimetre to centimetre-scale alternations of shale and very fine-grained sandstone. The latter tends to be more important westward in the study area. Monolithic shale or mudstone beds with sharp upper bounding surfaces and sharp or gradational lower bounding surfaces are prominent in the easternmost part of the study area,

and to a lesser degree westward at Seven Fountains and Waaipoort, where heterolithic iterations of this facies association are more common, as is the case also for the upper Weltevrede Formation. The lateral extent is seldom possible to ascertain from outcrop, however, two mudstone horizons in the lower Witpoort Formation were traced for several kilometres along strike, one at Rabbit Ridge (Chapter 7) and the other in the lowermost Witpoort Formation in the vicinity of Howison's Poort.

Mudrock facies are sometimes directly underlain by a thin bed or stringer of mudclast conglomerate, where the intraclasts seldom exceed pebble size and are well-rounded. Together with the sharp change from sandstone to mudrock this conglomerate is interpreted as representing a ravinement surface (Fig. 5.28). Other examples of FA 2c however, show conformability to the underlying sandstone strata, with a heterolithic transition zone (FA 2b). Gradation from underlying heterolithic intervals, which have been interpreted above as deposits of the fluvial-tidal transition zone, according to Walther's Law, implies that the mud-dominated strata accumulated in adjacent environments.

Interbeds of sandstone similar to those forming FA 2b, which are very thin to medium cross-stratified, horizontal, undulatory or ripple cross laminated, sharply based beds, occur sporadically (Figs 5.32 and 5.33). They differ from FA 2b in that mudrock is here the predominant lithology. These sandstone interbeds indicate high energy events that interrupted the quiescence of mud accumulation. Two potential sources for the sandstone sediment are envisaged; they could have originated by marine washover processes or by fluvial flood events, which may be distinguished by flow vectors. For example, Figure 5.33 illustrates a pair of thin cross-stratified and ripple-laminated sandstone beds contained within a 4–5 m thick black shale horizon at Waterloo Farm (WF2). The beds exhibit a west to west-northwest vector signal, which would be more consistent with a fluvial rather than marine source for the

sediment, according to the flow vector trends for the remainder of the study area (see 8.3 below).



Figure 5.33: The upper shale at Waterloo Farm exposes one of the better preserved examples of the estuarine facies assemblage (FA 2) from the study area (for a strat. log and interpretation see Figure 8.1F).

Concentrations of comminuted plant fossils are a near ubiquitous component of this facies association. They regularly occur in parallel or sub-parallel alignment. More complete plant remains are limited to specific exposures, and tend to favour the more monolithic, finer-grained mudrock, such as that exposed at Waterloo Farm and Coombs Hill. Plant remains form typical Late Devonian assemblages including arborescent and herbaceous or shrubby taxa as well as marine and non-marine algae (Chapter 2.2.2). Plant taphonomy varies markedly between and within horizons. Relatively complete remains, including metre-plus sized intact plants with multiple orders of branching occur in three mudstone horizons at Coombs Hill in association with mats of disarticulated plant and algal matter. At Waterloo Farm plants exhibit a similar taphonomic variety, but in finely laminated black shale facies,

appear to be more disarticulated and flattened than in the more thickly laminated facies (Gess 2011).



Figure 5.34: An example of Facies Association 2c from the study area (Waterloo Farm, section WF1b). The photograph shows the upper 40 cm of a 2 m thick upward-fining heterolithic unit. In its lower part, clay draped climbing ripple cross-laminae prograding north to northeast (landwards) pass upwards to undulatory and horizontal, very thinly interlaminated fine-grained sandstone and shale. Weakly developed, rhythmic laminasets from 0.5–3 cm thick show both fining and coarsening upwards, which do not show a clearly repetitive cyclicity. It is not determined whether these rhythmites represent varves, tidal cyclicities or otherwise. In the centre of the photograph, flaser and lenticular bedding with bidirectional ripple cross laminae (no palaeocurrent data recorded) pass upwards conformably to weakly laminated carbonaceous black shale at the top of the unit, which is sharply overlain by a thinner unit of similar facies.



Figure 5.35: Mudstone from Coombs Hill. Stratification in this rock is limited to horizontal alternating red and grey laminae. The red, oxidised layers are discontinuous, and are the product of secondary iron oxidation. They are subordinate in thickness to the grey layers. This stratification is interpreted as evidence for seasonal cyclicity.

Stratification of mudrock in the study area encompasses significant variation ranging from very thinly laminated to thickly laminated. These variations can form under similar conditions as evidenced by intergradations between them (Figs 5.34 and 5.35). Comparing the thin laminated black shale of Waterloo Farm MFL with the more crudely bedded black mudstone of the Blue Layer it is hypothesised that the rate of deposition is the main control. Evidently the Blue Layer with its more complete, three-dimensional plant remains was more rapidly deposited, whereas greater compression of plants in the MFL suggests slower deposition. Similar rapid burial is indicated by Coombs Hill plant remains, where individual three-dimensional plant specimens may occupy almost a centimetre of rock thickness perpendicular to bedding (Harris 2019). It is likely therefore muds would have accumulated in the estuaries at rates of at least several centimetres per annum (Fig. 5.35). Alternating cm-scale grey and red oxide laminae in the plant bearing beds at the locality may therefore represent seasonal cycles, reflecting geochemical shifts between iron oxidation and reduction (Hoosen 2019).

FA 2c is interpreted as having formed in the zone of turbidity maximum in estuaries, seaward of the fluvial-tidal transition. The estuarine turbidity maximum is a zone of elevated suspended sediment concentrations occupying the middle part of estuaries and deltas that arises as a result of estuarine circulation, or the up-estuary transport of sediment driven by density differences between saline and fresh waters (Uncles *et al.* 2006; Dalrymple and Choi 2007). Within this zone, muds can accumulate at rates of decimetres to metres per annum, as observed in modern estuaries (Shephard and Moore 1960; Reineck and Singh 1980; Dalrymple *et al.* 1992; Cowan *et al.* 1998). Rapid accumulation of muds is facilitated by flocculation, a process that is most prolific in brackish water, at salinities ranging from 1–10 per mil (seawater is about 35 per mil) (Dalrymple and Choi 2007). This turbidity maximum is a dynamic zone that can occupy a large region of an estuary, extending for up to several tens of km in modern estuaries, approaching the limit of salinity intrusion (Wolanski *et al.* 1995; Hughes *et al.* 1998; Uncles *et al.* 2006; Doxaran *et al.* 2009). Here, suspended sediment concentrations (SSCs) can reach several orders of magnitude higher than in adjacent environments (Dalrymple and Choi 2007). For example, the macrotidal Humber estuary produces a turbidity maximum zone up to 35 km long, particularly during periods of low fluvial discharge, reaching as little as 3 km seaward of the tidal limit, which is several tens of km from the mouth (Uncles *et al.* 2006). As with salinity gradients this zone moves up and down estuary in response to tidal and fluvial cycles and its properties are dependent on the interplay of these factors. SSCs can reach far greater levels in microtidal systems where circulation within the estuary is restricted (Hughes *et al.* 1988).

Deposition within the estuarine turbidity maximum zone is consistent with palaeontological evidence arising from strata at Waterloo Farm, Rabbit Ridge and Coombs Hill. Waterloo Farm black shale preserves marine and non-marine vertebrate taxa, monospecific epifaunal bivalves, eurypterid arthropods, terrestrial plants and marine and non-

marine algae (rhodophytes and charophytes) (Scholze and Gess 2017; Gess and Whitfield 2020). Rabbit Ridge preserves a monospecific lingulid brachiopod community associated with transported plant axes and arthrodire placoderm remains (this work, Harris and Gess 2022). Ichnofossils from FA2c comprising *Zoophycos*, *Spirophyton*, *Ptychoplasma*, *Planolites*, *Lockeia*, *Lingulichnus*, *Undichna* and piscean coprolites are likewise consistent with this depositional environment.

FA 2d– Outer estuary channels and sandbars

Certain sandstone strata resembling FA 2a were found to be inconsistent with fluvio-estuarine channels as described above after restoration of the flow vectors contained therein. The features of FA 2d are medium to thick lenticular and sigmoidal beforms, mud-draped foresets, irregular lamination, reactivation surfaces and silt or mud-rich toe sets with ripple cross-laminae. The facies assemblage thus resembles FA 2a in many respects, however, the sediment transport direction is opposite. FA 2d was only recognised in the lower Rooirand Member at Waaipoort, and there is no ichnological or palaeontological data to support its distinction.

An example is illustrated in Figure 5.36, which is a four metre thick heterolithic bedset comprising thick bedded, fine-grained, trough cross-stratified sandstone (St) with mud-draped foresets and muddy interbeds with flaser and lenticular ripples (Sr (b)). In this deposit, double mud drapes are locally developed. Cross-stratification is irregular, with muddier and sandier foresets developed in inclined laminasets of up to 40 cm thick separated by reactivation surfaces (Fig. 5.36C). The unit is underlain by FAs 2a and 2b. Based on the landwards (N to NE) flow vectors, mud-drapes and muddy interbeds, the unit is interpreted as reflecting deposition in flood-dominated tidal inlets (cf. Cheel and Leckie 1990). The laminasets could record variation in current energy during the bimonthly tidal cycle (Visser 1980; Roep 1991; Van den Berg *et al.* 2007).

Flow vectors provide the clearest means of separating fluvio-estuarine channel from tidal channel deposits with the presently available dataset. More prominent mud drapes, finer-grained sandstones, and interbedded muds containing lenticular ripples appear to be characteristics that could further distinguish the latter.

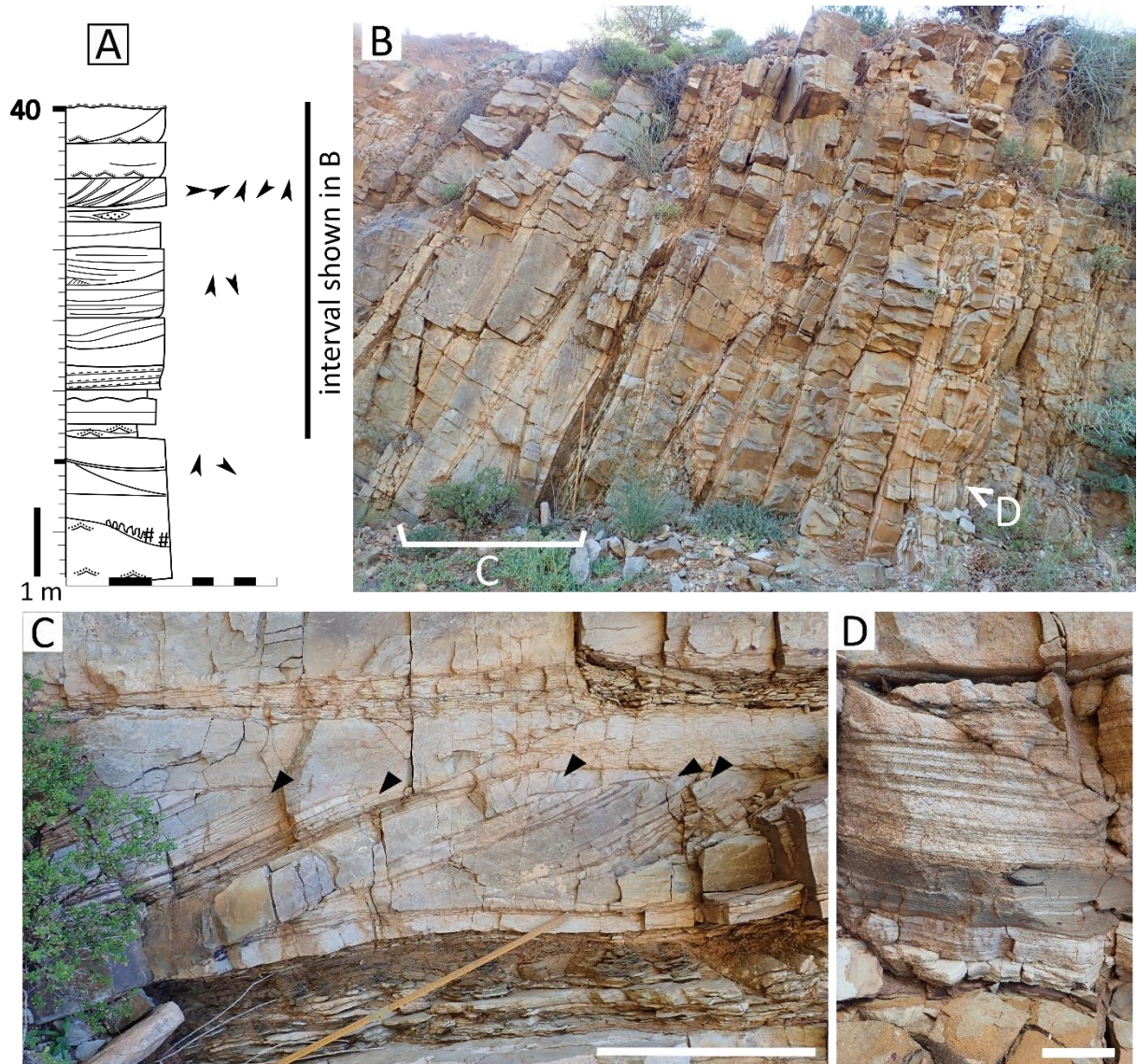


Figure 5.36: FA 2d in the lowermost Witpoort Formation at Waaipoort approximately 10 m above its basal contact. A) Stratigraphic log of the outcrop shown in B with flow vector orientations indicated by black arrows (WP 1, m. 33–40). B) Outcrop photograph indicating the locations of photograph enlargements C and D. C) Possible flood tidal inlet channel deposit. Black arrows indicate reactivation surfaces separating laminasets of mud-draped sand foresets (scale bar = 50 cm). D) Rhythmic laminated sandstone with mud drapes (scale bar = 5 cm).

CHAPTER 6 – ICHNOLOGY

6.1. Introduction

The interpretations gained by facies analysis in Chapter 5 predicted that many, if not all the ichnofossils from the study area were generated in estuarine settings. Ichnofossils provide a good test for this model because they are far more sensitive to environmental parameters such as salinity fluctuations than sediments (Buatois *et al.* 1997b). The estuarine environment is stressful for organisms due to its dynamism, with fluctuations in salinity, turbidity, oxygen levels and current energy coupled with high depositional rates that are reflected in the opportunistic bioturbation patterns therein (Ekdale 1985; Van den Berg *et al.* 2007; Buatois and Mángano 2011; Gingras *et al.* 2012).

Estuarine trace fossil assemblages fall into three broad categories representing the transition from freshwater to brackish and marine conditions. Tidally influenced, freshwater inner estuarine environments are inhabited by non-marine organisms, which rapidly disappear with elevations in salinity (Buatois *et al.* 1997b; Diez-Canseco *et al.* 2015; Shchepetkina *et al.* 2019). Estuarine central basins are generally characterised by brackish water ichnofaunas (Mángano and Buatois 2011), with high abundance, low diversity assemblages of infaunal marine traces that are generally reduced in size compared to their open-marine counterparts (MacEachern 1994; Buatois *et al.* 2005). Outer estuary deposits typically reflect higher energy deposition and rapid bedform migration which may inhibit burrowing. Greater marine influence in the outer estuary is reflected by a higher diversity of traces and behaviour types compared with the more restricted brackish water system, but is nonetheless depauperate compared with open-marine deposits (Mángano and Buatois 2011). The taxonomic composition of brackish water benthos and their resulting traces has changed with invertebrate evolution during the Phanerozoic, with a punctuated increase in burrowing depth and intensity (Buatois *et al.* 2005). Reviews of ichnology of the fluvial-tidal transition

zone indicate a shortage of studies from Devonian and older strata (Diez-Canseco *et al.* 2015; Shchepetkina *et al.* 2019).

Prior to this work, ichnology of the Witpoort Formation included three ichnogenera identified at Waterloo Farm (Gess and Hiller 1995a) and recognition of burrowed horizons elsewhere in the stratigraphic column (Brunsdon 2006; Vorster 2013). The rarity of trace and body fossils led Loock (1967, p. 104) to suggest that “the development of fauna was restricted” during deposition of the Witpoort Formation. The analysis in this chapter will demonstrate the markings of invertebrate and vertebrate activities at specific loci throughout the studied stratigraphic succession and attempt to systematically characterise them for a better understanding of the strata they inhabit.

Table 6-1: List of ichnotaxa from the study area, their stratigraphic ranges, tracemakers and ethology. The possible trace makers and behavioural categories are inferred based on the references cited in text.

Identification	Range	Possible tracemaker(s)	Behaviour
<i>Altichnus</i> spp.	lower Witpoort Fm	Worms?	Sessile suspension feeding and equilibrium, budding (asexual reproduction)
<i>Archaeonassa fossulata</i>	upper Weltevrede Fm to lower Witpoort Fm	Molluscs (gastropods?)	Surface locomotion and shallow ploughing, resting
<i>Arenicolites curvatus</i>	upper Rooirand Mbr	Worms or crustaceans	Dwelling and suspension feeding
<i>Cochlichnus anguineus</i>	upper Rooirand Mbr	Annelids	Infaunal grazing
<i>Cruziana problematica</i>	upper Rooirand Mbr	Notostrachans or trilobites	Infaunal grazing
<i>Gordia marina</i>	upper Weltevrede Fm	Worms, insects, gastropods and millipedes	Infaunal grazing
<i>Lingulichnus hamatus</i>	lower Rooirand Mbr	Lingulid brachiopods	Reburrowing
<i>Lingulichnus verticalis</i>	lower Rooirand Mbr	Lingulid brachiopods	Dwelling, suspension feeding

<i>Lockeia siliquaria</i>	upper Weltevrede Fm to upper Rooirand Mbr	Bivalves	Resting
<i>Nereites</i> ispp.	lower Rooirand Mbr	Worms (?)	Infaunal grazing
<i>Palaeophycus tubularis</i>	upper Weltevrede Fm to lower Witpoort Fm	Worms (?)	Open burrow
<i>Phycodes</i> cf. <i>palmatus</i>	lower to upper Rooirand Mbr	Annelid worms	Deposit feeding
<i>Planolites</i> ispp.	upper Weltevrede Fm to Skitterykloof Mbr	Polychaetes	Locomotion/grazing
<i>Psammichnites implexus</i>	lower Rooirand Mbr	Molluscs	Locomotion/grazing
<i>Ptychoplasma exelsum</i>	upper Weltevrede Fm and middle? Rooirand Mbr	Wedge-foot bivalves	Locomotion/grazing
<i>Rusophycus carbonarius</i>	upper Rooirand Mbr	Notostrachans or trilobites	Resting
<i>Scalichnus phiale</i>	Rooirand Mbr	Bivalves	Dwelling and suspension feeding
<i>Scalichnus lepusaures</i>	upper Weltevrede Fm to Skitterykloof Mbr	Bivalves	Dwelling and suspension feeding
<i>Selenichnites rossendalensis</i>	lower Rooirand Mbr	Limulids	Resting
<i>Undichna</i> ispp.	upper Rooirand Mbr	Fishes	Bottom swimming
<i>Zoophycos</i> cf. <i>brianteus</i>	lower Rooirand Mbr	Worms	Deposit feeding
<i>Spirophyton</i>	upper Weltevrede Fm to upper Rooirand Mbr	Worms	Deposit feeding
Plug-shaped burrows	upper Weltevrede Fm to upper Rooirand Mbr	Bivalves	Equilibrium or escape, dwelling
Diminutive plug-shaped burrows	upper Rooirand Mbr	?	Dwelling and equilibrium
Micaceous backfill structures	upper Rooirand Mbr	?	Shallow infaunal grazing
Compound trace 1	lower to upper Rooirand Mbr	Bivalves?	Infaunal grazing
Compound trace 2	upper Weltevrede Fm to upper Rooirand Mbr	Bivalves?	Infaunal grazing
Bromalites	Rooirand Mbr	Fishes and/or eurypterids	Excretion
Siphonichnidae indet.	Rooirand Mbr	Bivalves	Siphonal activities

6.2. Trace fossil descriptions

Ichnogenus *Altichnus* BROMLEY AND HANKEN 1991

Altichnus isp.

Altichnus isp. A

Fig. 6.1 A–C

Material– Tens of specimens at outcrop (not collected)

Horizon– Two levels in the lower Witpoort Formation (Rabbit Ridge) between 35 and 60 m above the base of the formation

Occurrence– Burrows are preserved as hollowed out moulds within the upper 7 cm of a thick ripple cross-laminated (Sr) bed in the lower section at Rabbit Ridge. Approximately 10 m of along-strike profile exposure provides 32 burrows, all of which were truncated at approximately the same level. A plan view of similarly-shaped burrows is encountered at a sandstone-siltstone contact in the middle section.

Description– Unlined j and y shaped burrows, up to 125 mm long, cylindrical in cross-section, widest at the top where sub-vertical (20–25 mm wide), tapering downwards to a curved sub-horizontal basal tube (2–3 mm wide at base). Burrow fill is structureless silty sand, softer than the matrix. Branching of the basal tube or abrupt tapering and widening of the burrow (Fig. 6.1A, B) may represent re-equilibration with the sediment surface during erosion and subsequent deposition.

Remarks– The type ichnospecies, *Altichnus foeyni*, from the Lower Cambrian of Norway includes passively filled, upward-branching conical burrows (Bromley and Hanken 1991).

Altichnus is considered to represent the domiciles of wormlike suspension feeders with vegetative budding (Bromley and Hanken 1991). The ichnogenus is typical of shallow-marine or intertidal settings (Bromley and Hanken 1991; Erdogan *et al.* 2004; Gaillard and Rachebeouf 2006; Penn-Clarke *et al.* 2019).

The fine-grained fill and attitude of the burrows distinguish the Rabbit Ridge material from the type material. The material is very similar to *Altichnus* isp. from the Lower Devonian of Bolivia (Gaillard and Racheboeuf 2006). In the Bolivian material, *Altichnus* isp. A includes vertically orientated traces not known to branch upwards, but which exhibit swellings interpreted as equilibrium structures. *Altichnus* isp. B includes inclined burrows with a curved, upward-branching basal tube, but equilibrium structures are unknown in the morphotype. The Rabbit Ridge specimens display a combination of characters resembling both these unnamed ichnospecies, but more closely resemble the Bolivian *Altichnus* sp. A. in size and attitude. Similarly shaped, branching examples occur in the Givetian strata of the Bokkeveld Group (Penn-Clarke *et al.* 2019).

Altichnus isp. B

Fig. 6.1 D–F

Material– **HP**: >10 specimens (uncollected)

Horizon– Lowermost Witpoort Formation, four metres above the Weltevrede Formation–Witpoort Formation contact.

Occurrence– Abundant burrows occur in the upper 15 cm of a structureless arenitic sandstone bed that is sharply overlain by carbonaceous siltstone.

Description– Vertical to subvertical, unlined burrow with swellings (maximum 14 mm diameter) and constrictions (minimum 3 mm diameter), branching upwards with an angle of about 50°. Straight to gently curved. Maximum observed length is 140 mm. Burrow fill is softer than the matrix. At its aperture the burrow is funnel shaped (Fig. 6.1E), tapering downwards to an elongate parallel sided basal tube.

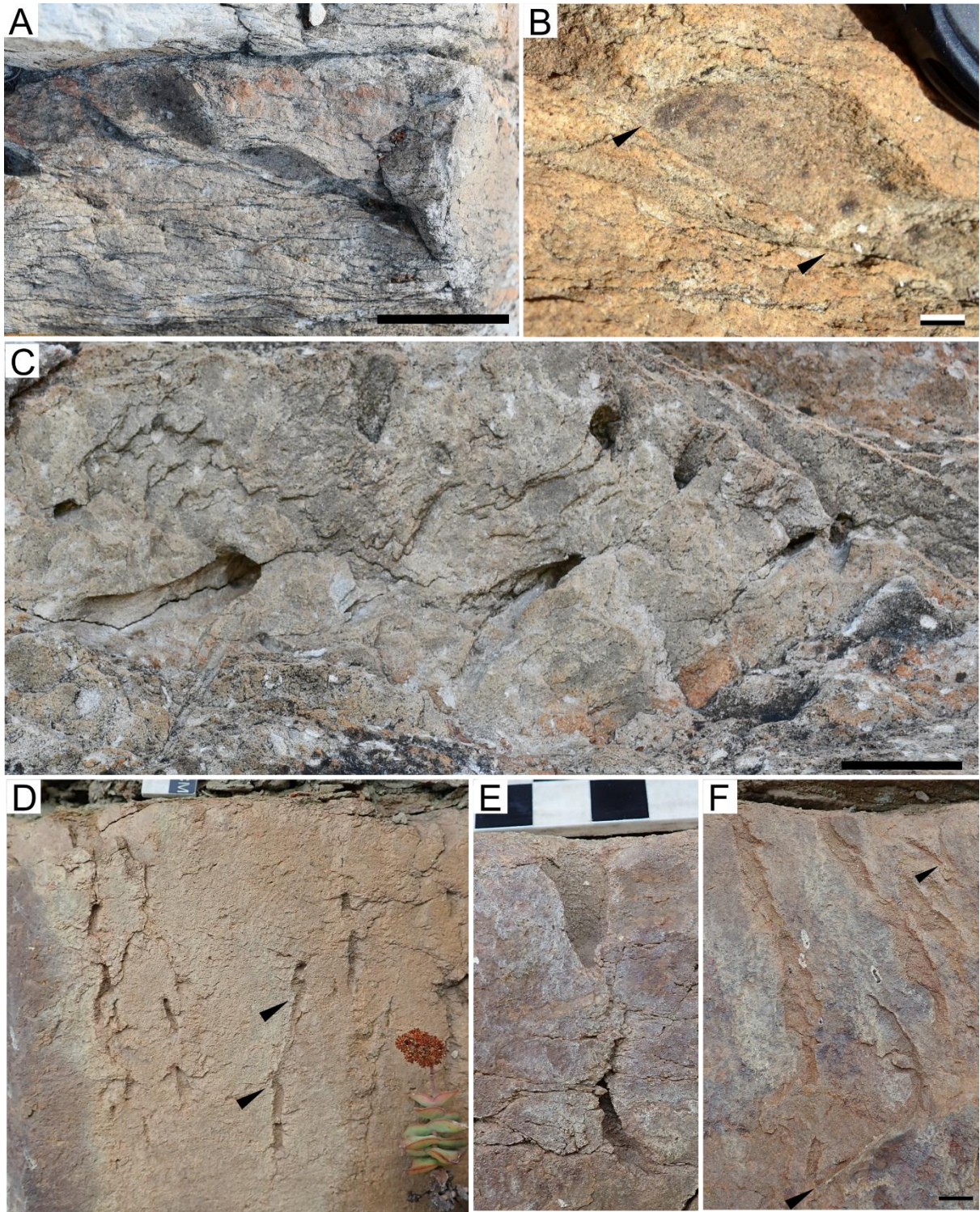


Figure 6.1: *Altichnus* in the lower Witpoort Formation (arrows indicate branching, all scale bars = 10 mm, except A, C = 50 mm and B = 5 mm): A–C: *Altichnus* (isp. A) from Rabbit Ridge; A) Profile view of branching burrow. B) Enlargement from A showing branching of the burrow indicated by black arrows. C) Multiple burrows viewed from underneath. D–F: *Altichnus* (isp. B) in the Lower Witpoort Formation at Howisons Poort. D) Profile view of burrows with apparent multiple levels of branching along their length. E) Funnel shaped aperture at surface. F) Burrows showing repeated pinchings and swellings and branches (arrowed).

Remarks– The burrows are similar to the type ichnospecies, *A. foyeni*, in shape with multiple swellings along the burrow length, upward-branching and the vertical attitude of the burrows (Bromley and Hanken 1991). They are distinguished from *A. foyeni* in having a burrow fill that is finer grained than the matrix, in which they are similar to *Altichnus* isp. B from the Devonian of Bolivia which also branches upwards (Gaillard and Racheboeuf 2006).

Ichnogenus ***Archaeonassa* FENTON AND FENTON 1937**

cf. *Archaeonassa fossulata* Fenton and Fenton 1937

Fig. 6.2 A–D, 6.3 A

Material– **HP**: AM 7658a (several specimens), AM 7661 (one specimen), uncollected material; **CH**: AM 7749 ai (one specimen), AM 7749 aii (at least four specimens), AM 7749 b (three specimens); **RR**: AM 7789 (7 specimens)

Horizon– Upper Weltevrede Formation to lower Witpoort Formation

Occurrence– A common trace fossil in thin-bedded heterolithic strata, occurring as epichnia on sandstone bed tops. In the Witpoort Formation the trace occurs in fining-upwards, thin-bedded sandstone with mud drapes (FA 2b) underlying mudstone (Coombs Hill and Rabbit Ridge). The Rabbit Ridge and Coombs Hill specimens are associated with paired and unpaired vertical siphonichnid traces. In the upper Weltevrede Formation the trace occurs in an upward-coarsening succession below a channel deposit (FA 2a), where abundant *Lockeia siliquaria* occur on the bottom surfaces of the same beds.

Description– Preserved on upper bed surfaces as a bedding-parallel furrow with lateral ridges, parallel-sided or pinching and swelling (diameter 6–20 mm; can vary by 25% along trace length) with a central furrow that occupies half, or less, of the total trace width. Two morphotypes are recognised, which co-occur and intergrade: Morphotype 1 is a U- or V-

shaped groove with irregular marginal levees that are variable in dimension, and may be equal or unequal, or absent. Morphotype 2 is a shallow flat-bottomed furrow with even, parallel marginal levees, sometimes showing transverse ornamentation and/or segmentation of the levees (segments at about 5 mm intervals). Some examples of the second morphotype appear to have been etched at surface before burial (Fig. 6.2D). Where preserved, levee segments are elliptical or lanceolate lobes that are parallel or oblique to the long axis of the trace. Traces of both morphotypes can terminate abruptly with a crater-like, actively-filled vertical structure, and this may mark a sharp change in tracemaker direction. This is evident on AM 7789, where several examples originate and terminate with shallow circular or irregular crater-like depressions slightly wider than the burrows (up to 20 mm) surrounded by pushup ridges. Clear gradation between morphotypes 1 and 2 is observed on two specimens on the slab (Figs 6.2 C, 6.3 B).

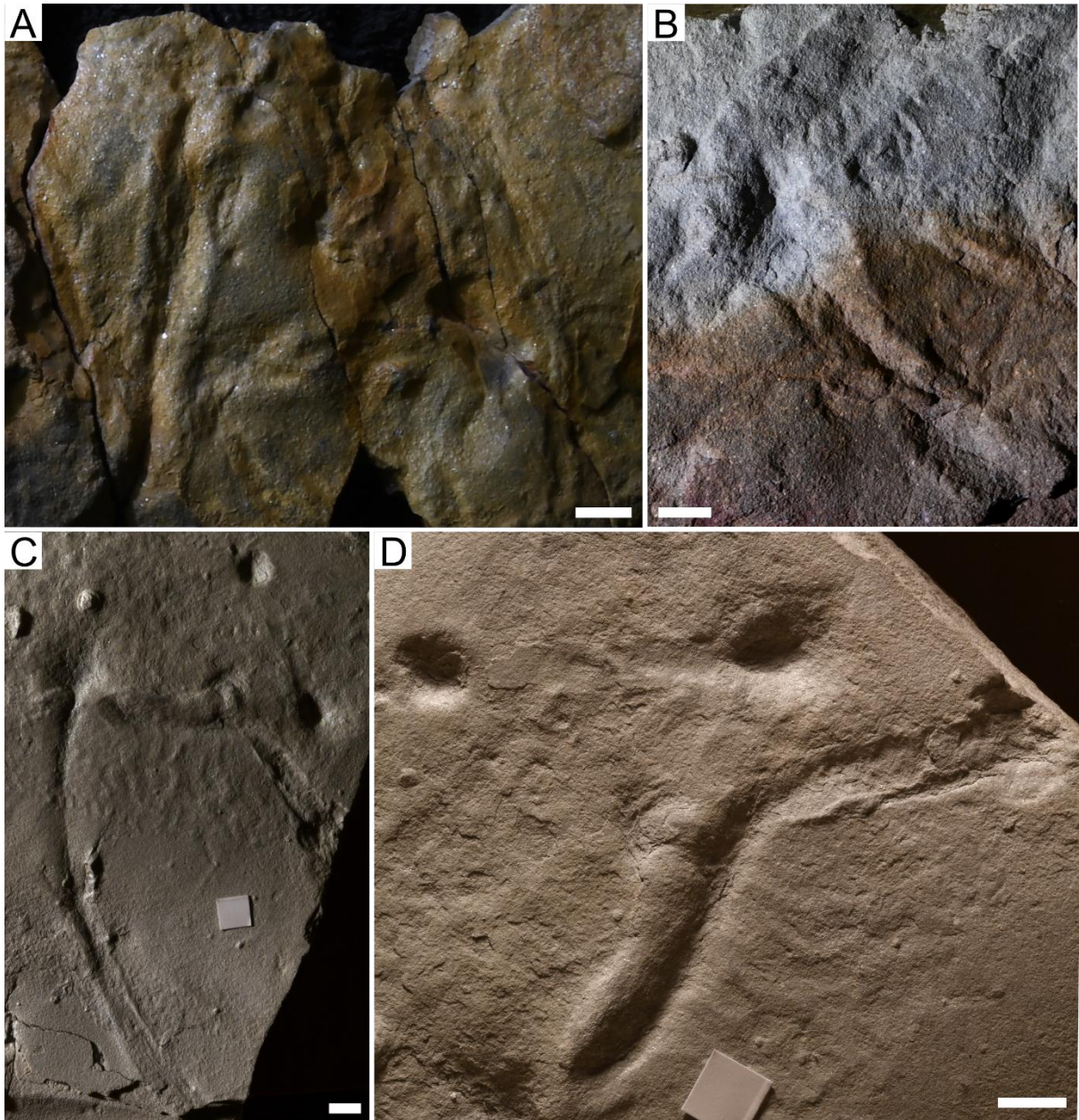


Figure 6.2: *Archaeonassa* cf. *fossulata* from the upper Weltevrede Formation (A) and Witpoort Formation (B–D) (all traces are epireliefs; scale bars represent 10 mm). A) Specimens from Howison's Poort (AM 7658), showing morphotype 1, interpreted as a shallow ploughing trail (left of photograph) and morphotype 2, interpreted as a surface trail (right of photograph), note the segmentation of the levees in morphotype 2. B) Specimen from Coombs Hill (AM 7748 j) showing slight segmentation of the marginal levees and a circular depression (resting trace) marking a change in direction. C) Traces on a wrinkle marked palaeosurface from Rabbit Ridge (AM 7789) showing gradation between morphotypes 1 and 2. All the traces on the slab originate and/or terminate with a circular resting structure. D) Enlargement from C showing the gradation between morphotypes 1 and 2. At the top of the photograph two shallow depressions are connected by a faint (washed out?) example of morphotype 2.

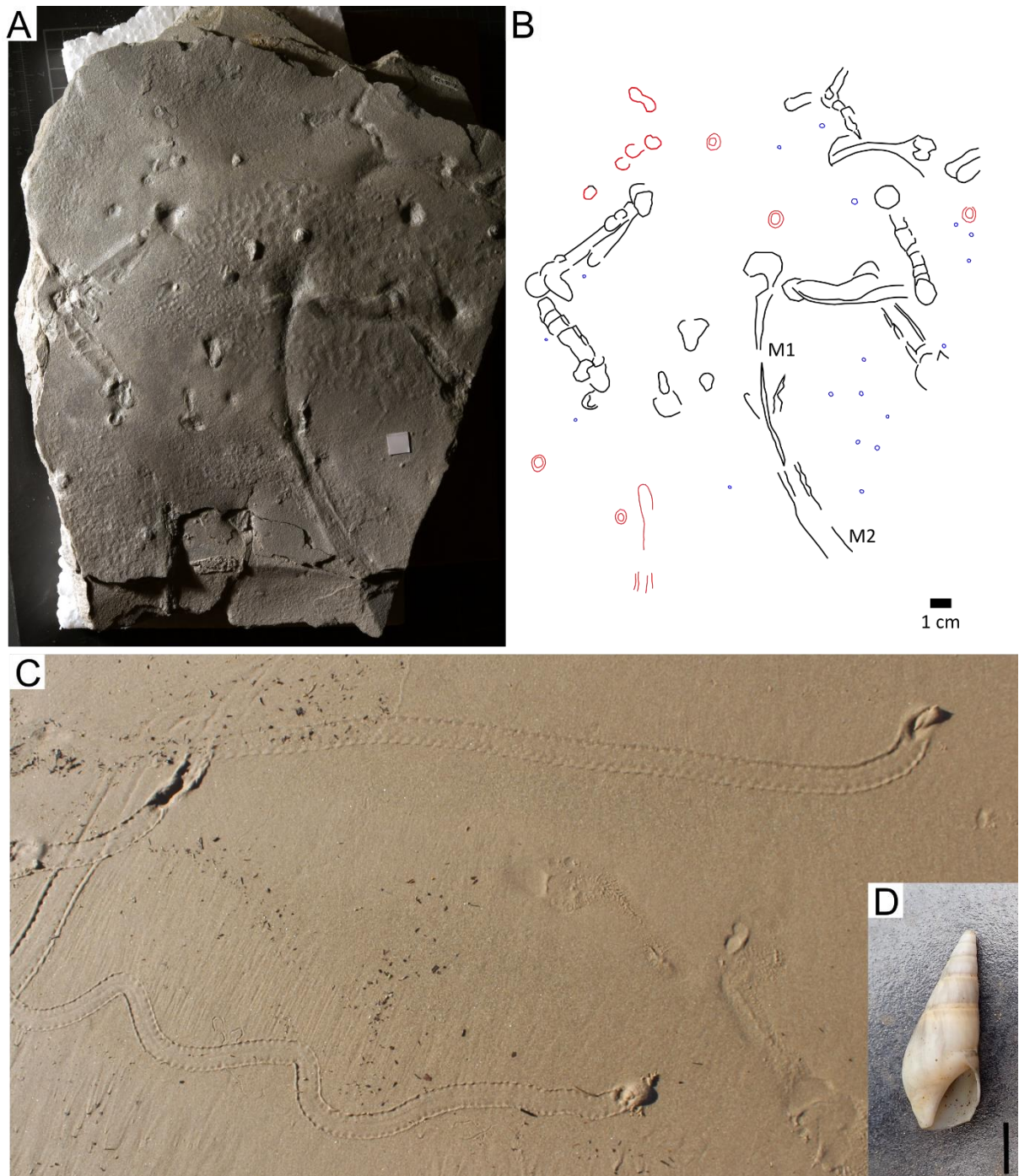


Figure 6.3: *Archaeonassa* from Rabbit Ridge compared with modern gastropod surface trails. A) *Archaeonassa* as a compound trace on a rippled and wrinkled palaeosurface (AM 7789), with interpretive drawing, B) showing gradation from morphotype 1 (M1) to morphotype 2 (M2). Indeterminate ‘Siphonichnidae’ originating from an underlying horizon are shown in red. C) Plough whelk (*Bullia digitalis*) trails exhibiting comparable gradations to those on AM 7789, with surface locomotion, shallow ploughing and resting. D) Plough whelk shell (scale bar = 1 cm).

Remarks– The ichnogenus *Archaeonassa* is originally designated for surface or shallow ploughing trails comprising a furrow with lateral ridges (Fenton and Fenton 1937). Segmentation of the lateral ridges in *Archaeonassa* is accounted for by rupturing of surface sediment with high tensile strength (Knox and Miller 1985; Jensen 2003). *Archaeonassa* may represent surface grazing (Buatois and Mángano 2011) or locomotion (Buckman 1994). The trace occurs in pre-Cambrian (Jensen 2003; Darroch *et al.* 2021) and Palaeozoic strata (Buckman 1994) but is uncommon in younger strata. *Archaeonassa* producers include gastropods (Buckman 1994; Buatois and Mángano 2011; Baucon 2021), bivalves (Pryor 1967; Chamberlain 1975), nereidid polychaetes (Yuanyuan *et al.* 2019) and, theoretically, arthropods (Yochelson and Fedonkin 1997). The trace has been considered characteristic of the intertidal zone (Buckman 1994), but also occurs in open-marine deposits (Bryant and Pickerill 1990; Buatois and Mángano 2011). The original interpretation of *Archaeonassa* was challenged by Yochelson and Fedonkin (1997) in their reinvestigation of the type material from the Cambrian Period of British Columbia, Canada. They contended that transverse striation is not visible on the holotype, although associated specimens from the same slab distinctively show this ornament (Fenton and Fenton 1937, Plate 1, Figure 2).

Morphotype 1, which occurs on what appears to be an algal-bound palaeosurface in the Rabbit Ridge material (Fig. 6.3A), is consistent with a surface grazing or locomotory trail. Gradation from morphotype 1 to 2 is interpreted as a shift from surface locomotion to shallow ploughing, consistent with previous observations of this ichnogenus (Jensen 2003). This behavioural shift may, in turn, be related to changing substrate consistency during exposure. Increased sharpness of trail preservation in the Cambrian mollusc trail *Climactichnites* has been linked to locomotion on exposed surfaces in the intertidal zone (Getty and Hagadorn 2009). A modern analogue for this behaviour is provided by gastropod trails in the swash zone along the southeastern coastline of South Africa (Fig. 6.3C). The

plough whelk (*Bullia digitalis*) inhabits the high energy swash zone, where it scavenges along the sandy beach. In saturated sediment, the animal may plough just beneath the surface producing a morphologically indistinct plough trail (cf. morphotype 2). Progressive drying of the sandy substrata initiates surface locomotion, forming a sharply-ornamented, flat-bottomed, segmented trail analogous to morphotype 1. The *Bullia* trail differs from *Archaeonassa* in having a marked central groove caused by dragging of the apex of the whelk's shell. Rapid desiccation of the surface renders locomotion progressively slower, whereupon it buries itself shallowly to rest. Such a resting trace may be analogous to the vortical structures that terminate *Archaeonassa* specimens from the present study, favouring a molluscan, rather than vermian, tracemaker. Periodic resting is recorded in a variety of intertidal molluscs, which migrate according to the water table (Chamberlain 1975; Mikulas *et al.* 2017).

The described traces are similar to *Archaeonassa fossulata*, the type species. Gradational changes in trace ornament and shifting from surface locomotion to shallow ploughing are evident in material attributed to the species from the Lower Carboniferous of Ireland (Buckman 1994). Although periodic resting traces are associated with modern *Archaeonassa*, this combination of behaviours is not recorded in *A. fossulata* and this ethological difference may require ichnospecific distinction for the Rabbit Ridge specimens.

Ichnogenus *Arenicolites* SALTER 1851

Arenicolites curvatus Goldring 1962

Fig. 6.4 A–D

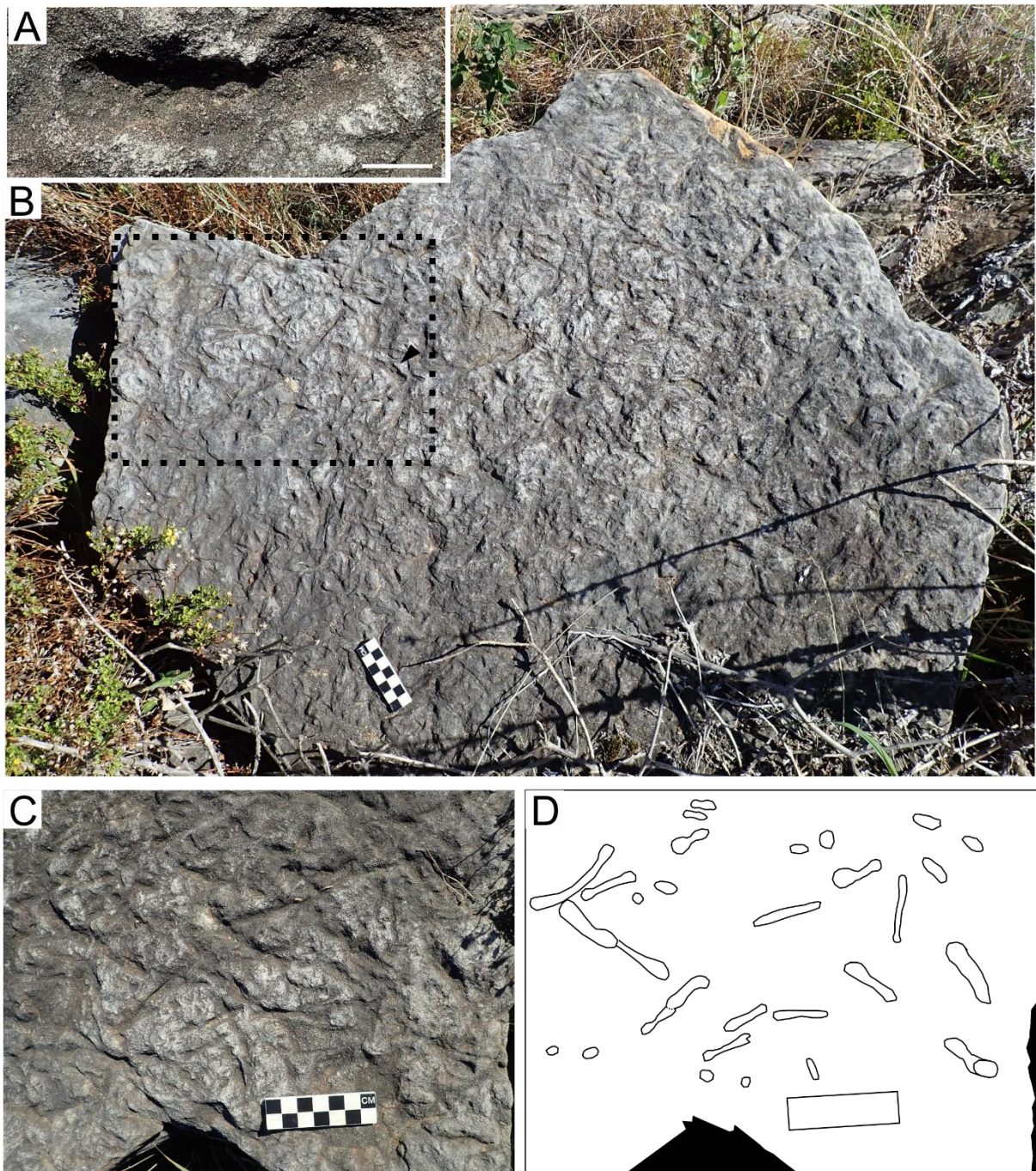


Figure 6.4: *Arenicolites curvatus* in the upper Witpoort Formation (scale increments represent 10 mm). A) Close up view of a specimen, note a slight dumbbell-shape and the presence of a prominent lining of light-coloured grains. B) General view of the slab containing the monospecific *Arenicolites* assemblage, stippled box shows the area shown in C (inverted) and white arrow indicates the trace enlarged in A. C) Enlargement from B with corresponding line drawing shown in D outlining some of the better preserved specimens.

Material– Tens of specimens from a large *ex situ* block situated at the top of WF 2c (not collected)

Horizon– Uppermost Rooirand Member

Occurrence– Highly abundant at the upper surface of a ripple cross-laminated medium grained sandstone greywacke bed. No other ichnotaxa are recorded at this horizon.

Description– U-shaped burrows preserved as negative epireliefs. Occurring as short, dumbbell-shaped horizontal burrows (up to 8 cm long, 5–12 mm wide), rounded and slightly flaring at both terminations, or as a pair of round to oval depressions of similar diameter that are both elongated in the same direction and inclined towards the centre. Where occurring as paired tubes, there is no evidence of sediment disturbance or spreite between the tubes. Burrow fill is structureless and softer than the sandy matrix. A prominent lining of mica grains about 1 mm thick is evident in some of the specimens (Fig. 6.4A). Depth of burrow penetration is not recorded and the traces likely represent the basal portions of u-tubes that extended into overlying sediment. Traces show two main orientations, one of which cross-cuts the other.

Remarks– Paired openings, which are in places connected by a tube that narrows towards the centre, suggest a simple u-tube. The absence of spreite places the material in *Arenicolites*, distinguishing the burrows from *Diplocraterion*. Although the type species of *Arenicolites* has been reinterpreted as a deformation structure in microbial mats, the ichnogenus remains in common use. *Arenicolites carbonaria* is considered the functional type ichnospecies (Goldring 1962; Rindsberg and Kopaska-Merkel 2005; Callow *et al.* 2011). The low angle of divergence of the arms of the u-tube is consistent with *A. curvatus*, which is a lined, arc-shaped tube from shallow-marine, possibly intertidal, strata in the Late Devonian of Devon (Goldring 1962). Comparable traces occur in shallow-marine strata in the Early Devonian of Antarctica (Bradshaw 2002), and in the lower Witteberg Group (pers. obs.). Knaust (2019)

attempted to synonymise this ichnospecies with *Tisooa habichi*, a vertical u-tube with a very high height/width ratio, but the shallow inclination of the arms of *A. curvatus* (40–60°) relative to bedding is not compatible with *Tisooa*. Bradshaw (2002) assigned shallowly inclined u-tubes to *Catenarichnus*, suggesting that *Arenicolites* be restricted to vertical u-tubes, however, the validity of *Catenarichnus* has been challenged (Buatois *et al.* 2017).

Arenicolites is very common in high-energy shallow- and marginal-marine deposits (including brackish water settings) (Goldring 1962; Bjerstedt 1987; Pollard *et al.* 1993; Buatois *et al.* 2015; Kappus and Lucas 2020) and is considered a typical member of the archetypal *Skolithos* ichnofacies (MacEachern *et al.* 2010). However, it is also known in ancient fluvial (Shillito and Davies 2017), lacustrine (Bromley and Asgaard 1979) and aeolian settings (Ekdale *et al.* 2007). Aquatic *Arenicolites* is usually considered to represent the permanent dwelling of a suspension feeder (Hakes 1976; Bromley and Asgaard 1979; Bjerstedt 1987). Based on neoichnological observations, potential tracemakers include worms and crustaceans (Goldring 1962; Bromley and Asgaard 1979).

Ichnogenus *Cochlichnus* Hitchcock 1858

Cochlichnus anguineus Hitchcock 1858

Fig. 6.5

Material– **WF:** AM 5265 (one specimen)

Horizon– Upper Rooirand Member

Occurrence– Epichnial relief in thin-bedded sandstone underlying the Main Fish Lens (MFL) black shale at Waterloo Farm. Rugose bioturbation on the same bedding surface might represent a poorly preserved example of ‘compound trace 1’. *Cruziana problematica*, *Rusophycus carbonarius* and *Phycodes palmatus*, ‘plug-shaped burrows’ and ‘compound trace 1’ occur on different slabs from a similar stratigraphic horizon at the locality.

Description– Sinusoidal trace in convex epirelief, 130 mm long, with regular meanders having a wavelength of about 25 mm. The trace is horizontal, but slight undulations in a vertical plane result in variations in apparent thickness, in places thinning or disappearing from the bedding plane for a short distance, particularly between the meanders, widening slightly at the turns, but is otherwise roughly parallel-sided (about 10 mm at widest point). The trace is sharp walled, lacks a visible lining, and comprises structureless sediment that is similar in composition to the matrix, with a slightly harder texture.

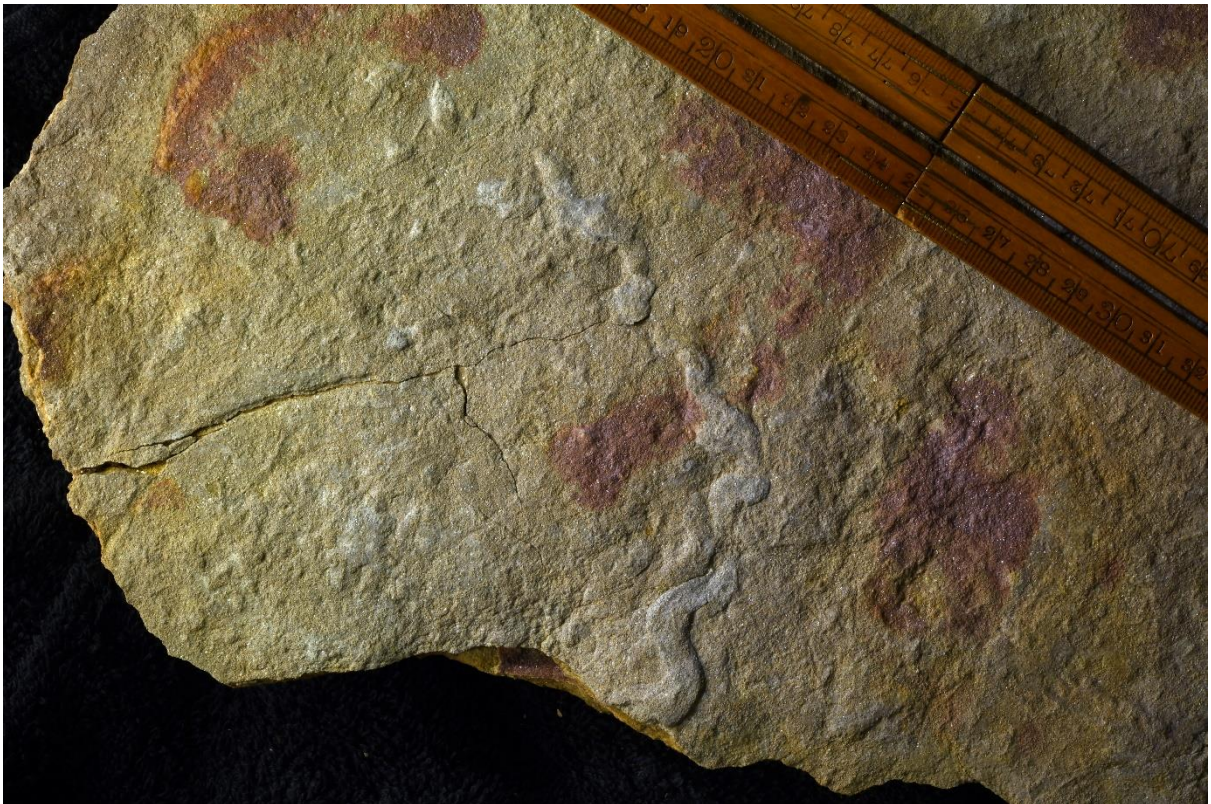


Figure 6.5: *Cochlichnus anguineus* from Waterloo Farm (upper Rooirand Member) in the Witpoort Formation (scale increments in cm).

Remarks– This specimen was previously illustrated and interpreted as a helical burrow possibly made by an annelid (Gess and Hiller 1995a). The trace is distinguished from the corkscrew-shaped burrow *Helicodromites*, which occurs as a series of oblique tubes representing successive whorls of a helix. Rather it has a sinusoidal form consistent with the

ichnogenus *Cochlichnus* (Uchman *et al.* 2009; Bordy *et al.* 2011), but it is unusually large. Similarly large *Cochlichnus* are known from Palaeozoic nonmarine, lake margin settings (Buatois *et al.* 1997a). The specimen is referable to *C. anguineus*, by the lack of annulations or lateral markings (Buatois and Mángano 1993). *Cochlichnus* is known from the Ediacaran to Recent and occurs in a wide variety of marine, marginal-marine and nonmarine environments (reviewed by Buatois *et al.* 1997a). The described specimen probably represents a shallow tier burrow combining locomotion and feeding (pascichnia). Possible tracemakers include nematode and annelid worms and some insects such as dipteran larvae (Buatois *et al.* 1997a; Uchman *et al.* 2009). The age of the strata probably excludes dipterans (Buatois *et al.* 1997a), and the large size of the traces would favour annelid rather than nematode worms, in agreement with the tracemaker suggested by Gess and Hiller (1995a).

Ichnogenus *Cruziana* D'ORBIGNY 1842

Cruziana problematica (Schindewolf 1921)

Fig. 6.6 A–D

Material– WF: AM 7625 (10's of specimens), AM 7626 (10's of specimens), AM 7610 (3 specimens), AM 7611 (8 specimens), AM 7609 (12 specimens), AM 5402 (c, e, f, j, l, n, v, w, y, z) (10's of specimens), AM 5403 (10's of specimens)

Horizon– Upper Rooirand Member at Waterloo Farm

Occurrence– Abundant on several bedding surfaces in a 1 m thick, thin-bedded heterolithic sandstone unit underlying the MFL black shale facies at Waterloo Farm. Traces exploit the bases of thin sandstone beds with wavy laminated and structureless immature greywacke at the interface with mudstone interbeds. The traces occur in sub-parallel alignment and are gradational to *Rusophycus carbonarius*. The latter is only absent from two slabs on which *Cruziana* are recorded. Many of the slabs contain only these ichnospecies,

however, less common associated ichnotaxa include ‘plug-shaped burrows’, ‘compound trace 1’ and *Phycodes* cf. *palmatus*.

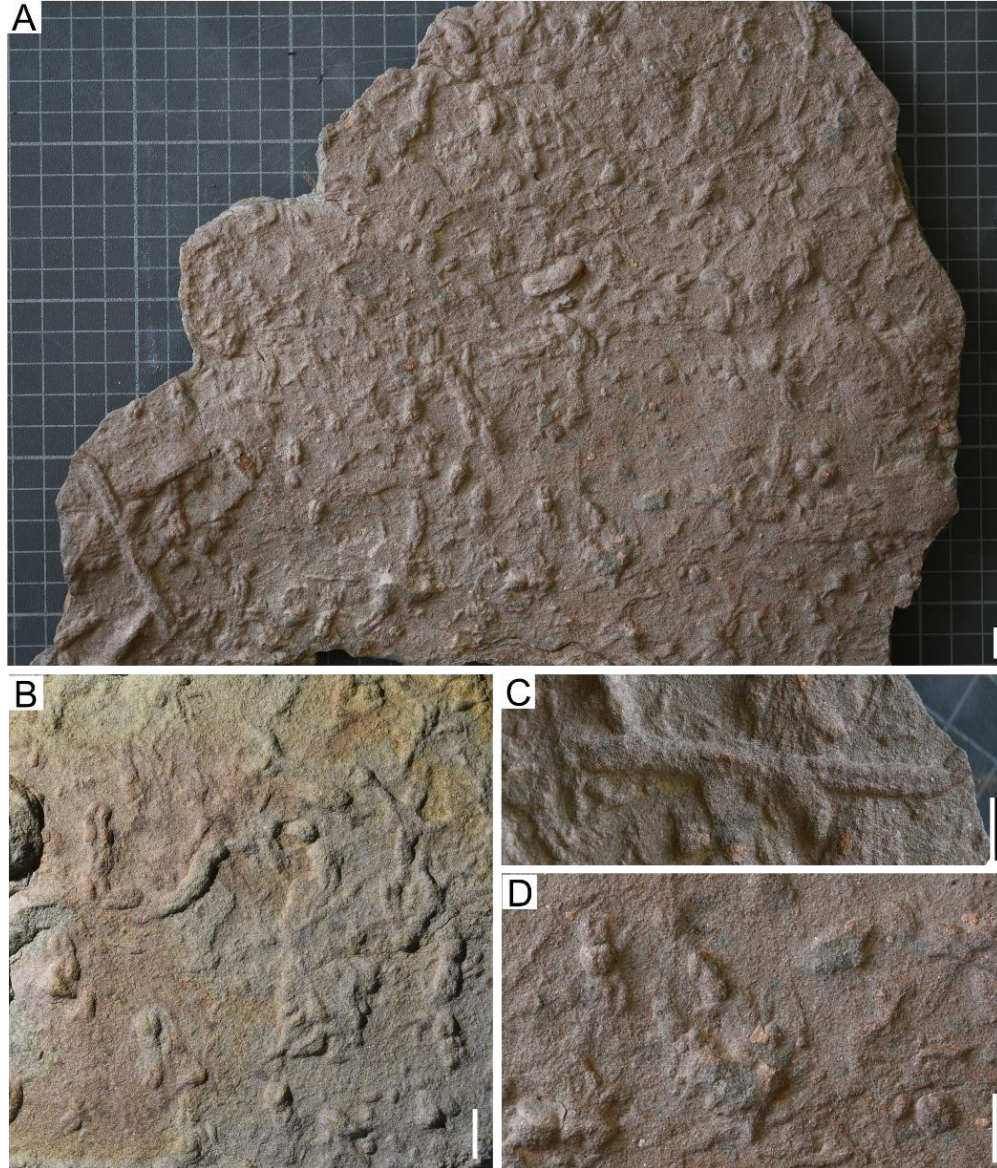


Figure 6.6: *Cruziana problematica* and *Rusophycos carbonarius* from the Witpoort Formation (all scale bars = 1 cm; all specimens in positive hyporelief). A) Slab with numerous specimens in sub-parallel alignment (AM 7625). B) Intergradation between *Cruziana* and *Rusophycus* (uncatalogued specimen). C) *Cruziana problematica* enlarged from bottom left of slab in A. D) *Rusophycus*, isolated and serially aligned, enlargement from photograph A, lower middle of slab.

Description– Bilobed, bedding-parallel trails, straight to gently curved, seldom exhibiting continuous meanders, up to 120 mm long, but usually less than 30 mm long, parallel-sided, 3–7 mm wide, comprising a pair of parallel hypichnial ridges less than 2 mm high, with a shallow median furrow (~1 mm wide). Trace is identical in composition to the host rock. Ridges occasionally have faint transverse ornamentation (Fig. 6.6C) as observed on specimens from AM 7627 and AM 7625. The traces subside at both ends of the trail where welded to the overlying bed. Many of the *Cruziana* comprise series of superimposed *Rusophycus*, with slight changes in orientation at short intervals of about 5 mm. This intermediate form is more common than straight *Cruziana* at this horizon.

Remarks– This material was identified as *Cruziana* by Gess and Hiller (1995a) and subsequently recognised as the *Isopodichnus* type (Almond 1998). The predominantly horizontal, bilobed traces are consistent with *Cruziana* or *Didymaulichnus*, but are distinguished from the latter by the faint ornamentation occasionally decorating the lobes (Keighley and Pickerill 1996; Buatois *et al.* 2017). The small size and subtle transverse ornament places the traces with *Cruziana problematica* (Keighley and Pickerill 1996). This ichnospecies is most common in continental and intertidal or estuarine settings, and less common in shallow-marine strata (Keighley and Pickerill 1996; Trewin 1976; Bromley and Asgaard, 1979; Mángano *et al.* 2002b; Mángano *et al.* 2023). Diminutive *Cruziana* from continental settings were previously referred to ‘*Isopodichnus problematicus*’ (Trewin 1976; Pollard 1985), which has been synonymised with *Cruziana* (Seilacher 1970; Bromley and Asgaard 1979; Keighley and Pickerill 1996). The traces record shallow ploughing of an arthropod beneath a thin veneer of sand (Bromley and Asgaard 1979). *Cruziana* is more often ascribed to the combined locomotion and deposit feeding by trilobites (Buatois and Mángano 2011), however, *Cruziana problematica* are likely produced by notostrachan (or similar)

crustaceans (Trewin 1976; Pollard, 1985; Schlirf *et al.* 2001). Trilobites are not excluded as possible tracemakers for *C. problematica* (Mángano *et al.* 2023).

Ichnogenus *Gordia* EMMONS 1844

Gordia marina Emmons 1844

Fig. 6.7 A, B

Material– **HP:** AM 7671

Occurrence– Hypichnia from the sandstone–mudstone interface in heterolithic, thin-bedded sandstone forming an upward-coarsening succession underlying a channel deposit (FA 2a). *Lockeia siliquaria* is abundant in the underlying and overlying beds.

Horizon– Upper Weltevrede Formation (approx. 50 m below the top of Fm) at Howison’s Poort

Description– Hypichnial, bedding-parallel, cylindrical structure, 2.5–5 mm in diameter, forming self-crossing loops (Fig. 6.7A). The burrow is a slightly flattened cylinder (Fig. 6.7B), massively filled with a similar composition to the host rock and with no evidence of a lining. A slight transverse ornament is observed on parts of the surface of the burrow.

Remarks– *Gordia* is a horizontal burrow characterised by looping and self-crossing (Buatois *et al.* 1998b; Keighley and Pickerill 1997; Uchman *et al.* 2009). It occurs in a variety of marine and intertidal (Buatois *et al.* 1998b; Hofmann *et al.* 2012) and non-marine settings (Keighley and Pickerill 1997; Uchman *et al.* 2009) and is regarded as a facies-crossing trace fossil. Possible tracemakers include polychaetes, nematodes, crustaceans, gastropods, insects and millipedes (Uchman *et al.* 2009; Hofmann *et al.* 2012; Wang and Rahman 2023). The present material lacks the sharp turns of *G. indianensi* (Buatois *et al.* 1998b) and it does not have the annulations characteristic of *G. nodosa* (Pickerill and Peel

1990). The smooth surface and open loops are consistent with *G. marina* (Keighley and Pickerill 1997).

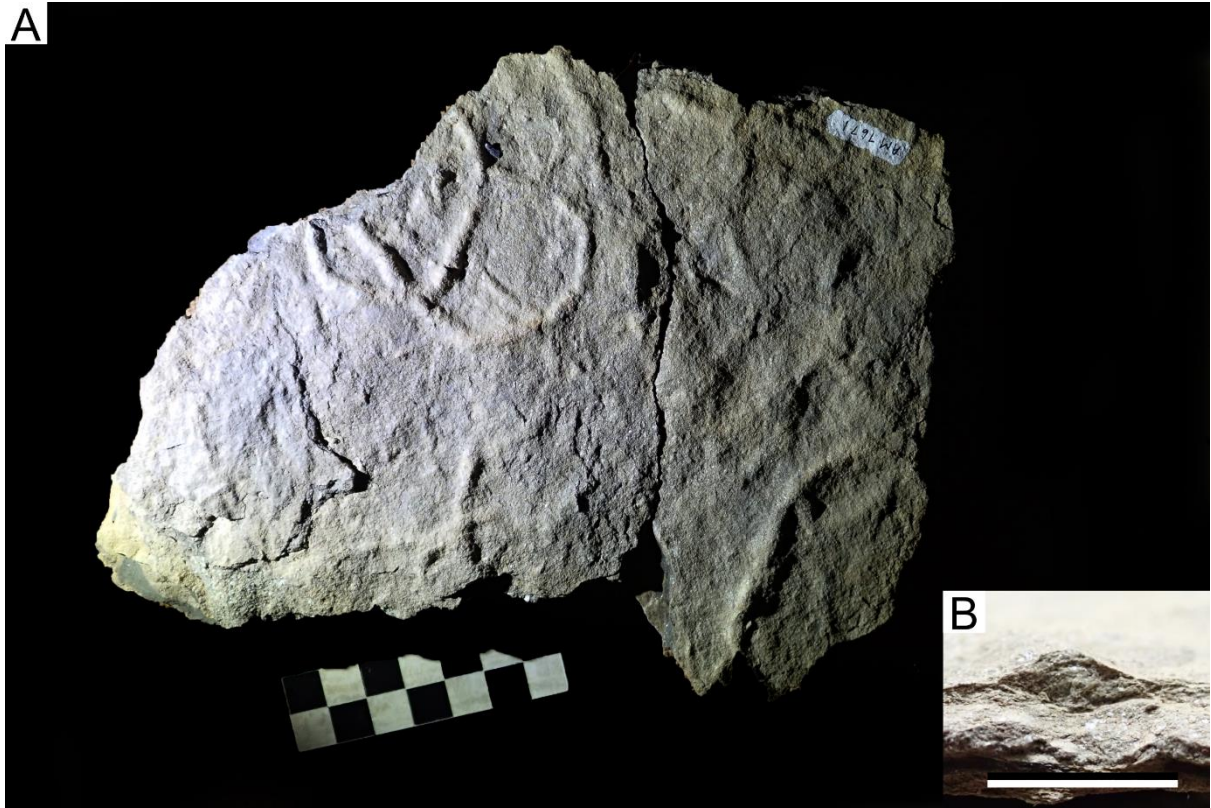


Figure 6.7: *Gordia marina* (upper Weltevrede Formation, AM 7671) (scale increments are 10 mm).

A) Bottom surface of the slab. **B)** Profile view of one burrow.

Ichnogenus *Lingulichnus* HAKES 1976

Discussion– Three ichnospecies of *Lingulichnus* are recognised as reflecting the various infaunal activities of lingulid brachiopods, including dwelling and reburrowing (Zonneveld and Pemberton 2003). *L. verticalis*, the type ichnospecies, is a vertical trace with an oval upper shaft above a narrower, circular basal stem. The upper shaft represents the portion of the burrow in which the shell moved upwards and downwards in response to environmental stimuli, whereas the lower stem contained the pedicle, which was anchored at the bottom. A thick mucous lining would have consolidated the burrow walls (Hakes 1976; Emig 1997). *L.*

inclinatus is distinguished from the type ichnospecies based on an orientation that is oblique to bedding (Zonneveld and Pemberton 2003). *L. hamatus* is a U- or J-shaped structure that records reburrowing by the brachiopod following exhumation (Zonneveld and Pemberton 2003). *Lingulichnus verticalis* ranges from the lower Cambrian (Pemberton and Kobluk 1978) to the present day. It is common in Palaeozoic marginal and shallow-marine environments and is also known in deep-marine strata (Paz *et al.* 2023), mimicking the body fossil record of lingulid brachiopods (Emig 1997).

Occurrence– Two *Lingulichnus* ichnospecies are recorded at one locality, a ~10 m thick carbonaceous mudstone horizon and associated sandstones at Rabbit Ridge. Burrows occur in immature sandstones underlying and interbedded with the mudstone, as well as within the mudstone itself. Burrows are associated with superabundant lingulid brachiopods (the inferred tracemakers) which occur *in situ* and scattered on bedding surfaces. Associated ichnofauna include ‘compound traces 1 and 2’, *Psammichnites implexus*, *Nereites* isp., *Phycodes palmatus*, and *Zoophycos* isp.

Lingulichnus verticalis Hakes 1976

Fig. 6.8 A–E

Material– **RR:** AM 7787a + b; AM 7786 (a+ b) (four specimens); AM 7791 a; AM 7812; AM 7806

Horizon– Lower Rooirand Member, Witpoort Formation

Description– Downwards tapering, thick-walled, vertical to inclined, straight burrows, filled with sediment that is distinct in colour and/or texture from the matrix, but similar in grain size. In plan view, burrows are fusiform or oval, with a long diameter of 6–22 mm. Diameter of the long axis is up to double that of the short axis. One burrow in plan view shows a central oxidised cylinder that might represent a pedicle trace (Fig. 6.8B). Burrow walls are 1–2 mm thick. Elliptical and smaller cylindrical forms occur in the same beds. In

profile the elliptical form has irregular margins, whereas the cylindrical form is even walled. Burrows range from vertical to inclined at 30° to bedding, and this range of orientation is observed on one slab (AM 7786; Fig. 6.8A). Maximum observed length of burrows is 83 mm, although complete forms may have been considerably longer.

Remarks– Size and shape, thick lining and co-occurrence with the inferred tracemaker supports the assignment to *Lingulichnus*. Co-occurrence of vertically and obliquely disposed burrows does not justify separating *L. inclinatus* from *L. verticalis* in this material, which is therefore included in the type ichnospecies, *L. verticalis*, regardless of angle of inclination. Associated *in situ* brachiopods from the same deposit include both vertical and oblique specimens (Chapter 7). The wider, fusiform burrow is the chamber inhabited by the lingulid. Underlying this, a narrower, more cylindrical structure contained the pedicle, the ‘basal stem’ (Zonneveld and Pemberton 2003). Oxidation of the burrow fill may indicate that some organic matter remained in the burrow, such as the animal body remains or abundant mucous that normally lines lingulid burrows.

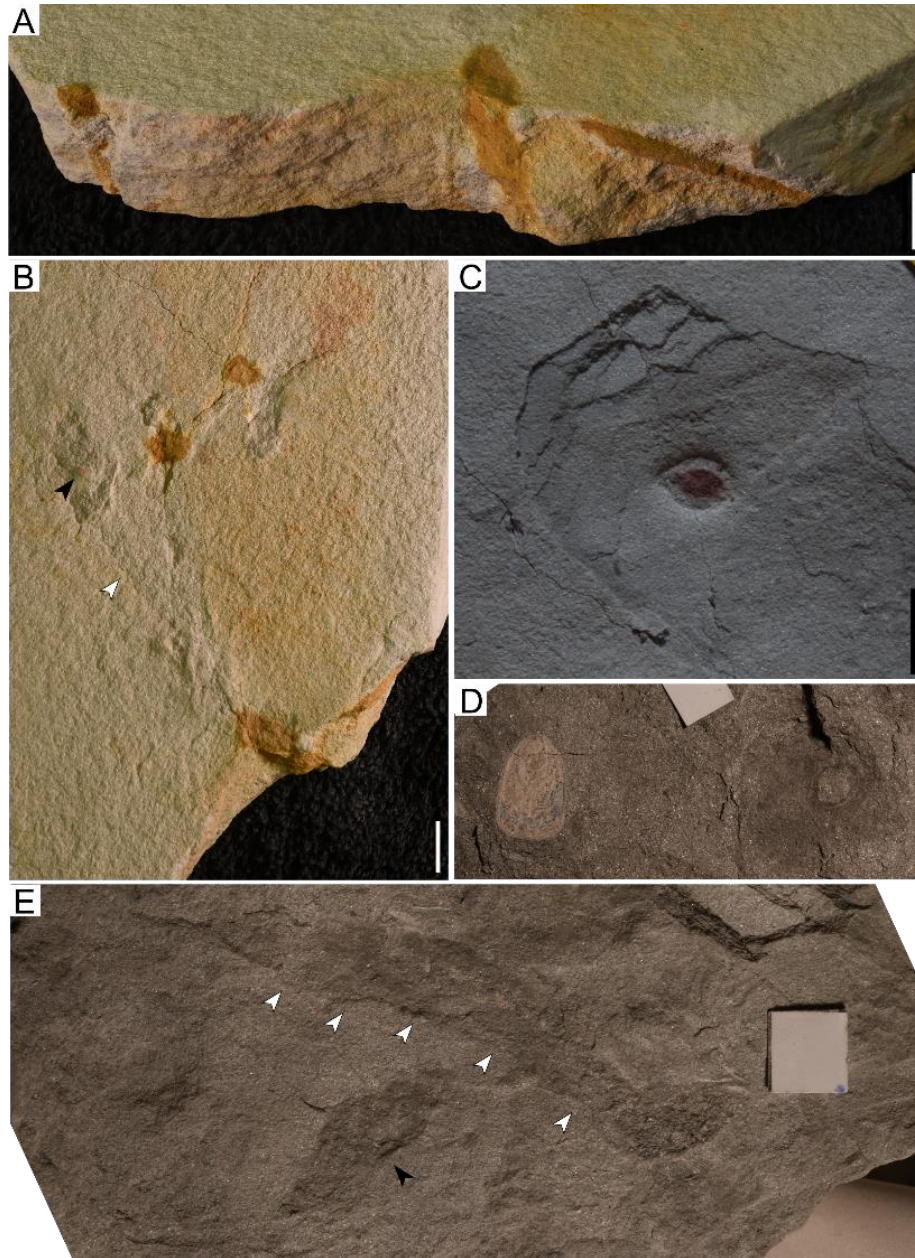


Figure 6.8: *Lingulichnus* traces from the lower Witpoort Formation Rabbit Ridge locality (all scale bars and squares indicate a centimetre): A) *L. verticalis*, oblique view (AM 7786). B) Bedding plane expression of *L. verticalis* and *L. hamatus* (AM 7786). Black arrow indicates a pedicle trace central to a vertical burrow. A white arrow indicates subtle sediment disturbance along the bedding plane (*L. hamatus*) suggesting the path followed by the tracemaker. C–D are interpreted as basal stem traces. C) Thick-walled *L. verticalis* showing internal oxidation (AM 7791a). D) *L. verticalis* associated with a disarticulated lingulid brachiopod shell on the same bedding surface (AM 7812). E) A possible lingulid reburrowing structure (*L. hamatus*) indicated with white arrows and *L. verticalis* shown with a black arrow (AM 7813).

Lingulichnus hamatus Zonneveld and Pemberton 2003

Fig. 6.8 B, E

Material– RR: AM 7786 (a + b); AM 7806

Occurrence– One specimen comes from parallel laminated fine-grained sandstone interbeds in the mudstone unit, and one specimen occurs in mudstone, associated with superabundant lingulid brachiopods. Both specimens are associated on the same bedding horizons with *L. verticalis*.

Description– A linear, bedding-parallel zone of sediment disturbance with indistinct, irregular margins, terminating in *L. verticalis*. These burrows are the same colour or slightly darker in colour than the matrix.

Remarks– These irregular bedding-parallel disturbances are placed in *L. hamatus* (Zonneveld and Pemberton 2003), because of the connection with *L. verticalis*. They are interpreted as the reburrowing traces of lingulid brachiopods. Reburrowing in extant *Lingula* was studied experimentally (Hammen and Lum 1977; Hammond 1983). This is a strategy for reestablishment after erosional exhumation or detachment from the pedicle. Escape during rapid deposition can induce detachment because the pedicle is anchored at the bottom of the burrow. Reburrowing capabilities are variable amongst *Lingula* species; *L. anatina* was not able to reestablish itself after experimental removal from the sediment (Hammond 1983), whereas *L. reevii* can reburrow following exhumation, with or without its pedicle (Hammen and Lum 1977). The pedicle may be regenerated after reburrowing (Hammen and Lum 1977).

Ichnogenus *Lockeia* JAMES 1879*Lockeia siliquaria* James 1879

Fig. 6.9 A–E

Occurrence– A common trace in the study area, occurring on the bases of sandstone and siltstone beds. In the upper Weltevrede Formation the trace is superabundant at specific horizons (BPBI 3–4), covering nearly the entire bottom surfaces of some sandstone beds (Fig. 6.9B, D and E). In the Witpoort Formation the trace is usually recorded in much lower densities as isolated occurrences in siltstone or sandstone. *Lockeia* is associated on the same bedding surfaces as *Ptychoplasma*, ‘compound trace 1’, ‘plug-shaped burrows’ and *Planolites*. *L. siliquaria*, in some cases, intergrades with *Ptychoplasma* in the study area (Fig. 6.9A and E).

Material– **WF**: AM 7626 (at least three specimens), **HP**: AM 7672 (12 specimens), AM 7683 (10’s of specimens), uncollected material from sections WF2c, WP and SF.

Horizon– Upper Weltevrede Formation to uppermost Witpoort Formation (Skitterykloof Member)

Description– Bilaterally symmetrical, amygdaloid to ovoid hypichnial protruberances, with pointed or rounded terminations, a smooth surface sometimes with a median keel. In the studied sample the traces are 7–20 mm long, 4–13 mm wide, and up to 2 mm deep.

Remarks– *Lockeia* is considered to represent the resting traces (cubichnia) of bivalves (Schlirf 2005). It is a widespread trace, common throughout the Phanerozoic. It is a facies-crossing ichnogenus (Goldring *et al.* 2005) that appeared in non-marine deposits at least since the Devonian Period (Buatois *et al.* 1998a), and is particularly prolific in Palaeozoic marginal-marine deposits (Buatois *et al.* 2005). The almond-shaped to oval trace is thought to be formed by a wedge-like foot of a bivalve (Seilacher and Seilacher 1994; Schlirf 2005; Seilacher 2007), however, interconnection with *Protovirgularia* shows that cleft foot bivalves

can also produce *Lockeia* (Ekdale and Bromley 2001; López Cabrera *et al.* 2023).

Conchostracan crustaceans can also hypothetically form similar traces (Bromley and Asgaard 1979). *Lockeia* commonly forms parts of compound trace fossils involving locomotion, resting and feeding (Ekdale and Bromley 2001; Knaust 2004; López Cabrera *et al.* 2023), and may also be associated with vertical equilibrium (escape) structures (Archer and Maples 1984). *L. siliquaria* has been separated from *L. amygdaloides* in having a more elongated shape (Knaust 2004), however, the two ichnospecies are alternatively considered synonymous Mángano *et al.* (2002).

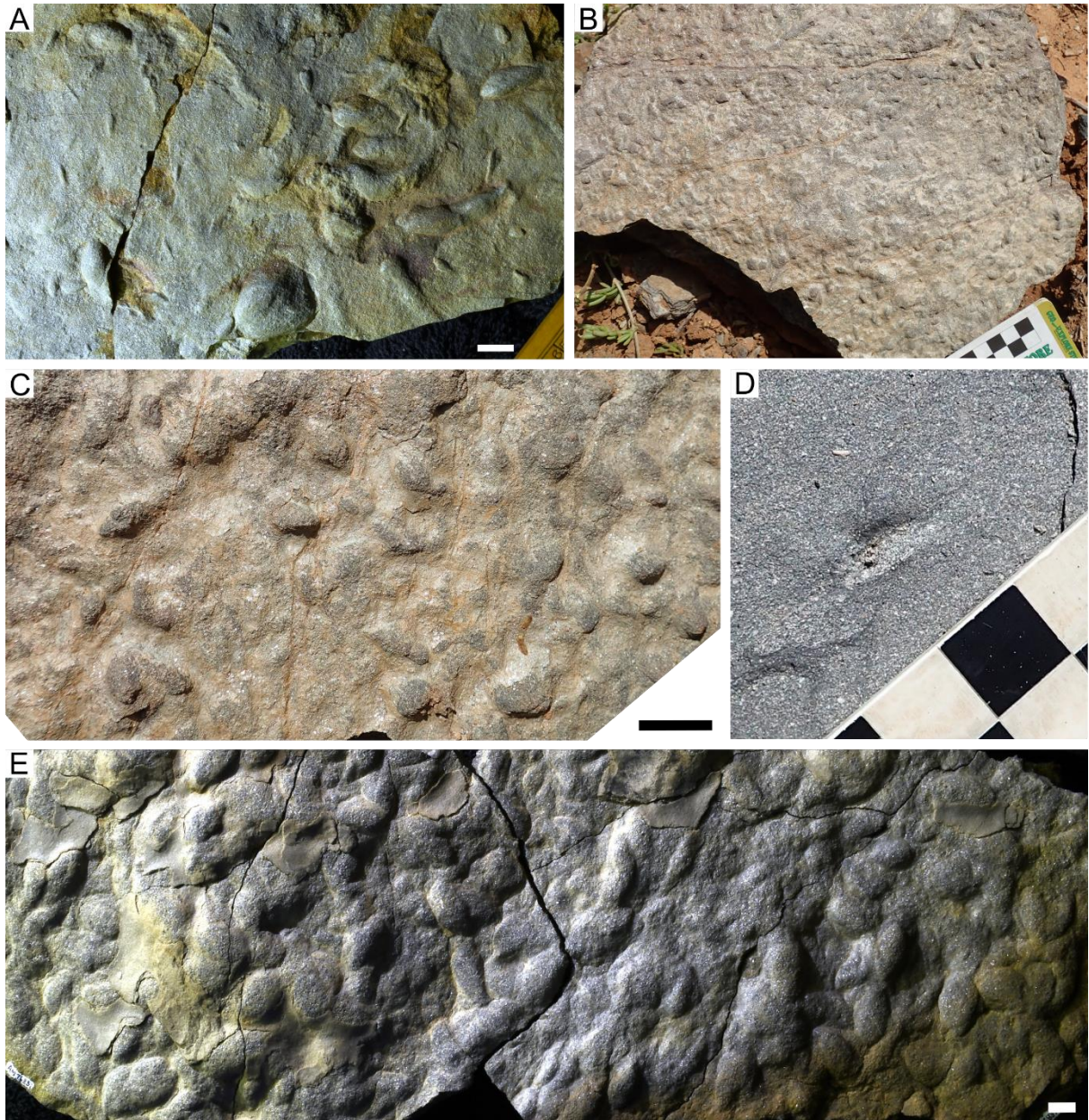


Figure 6.9: *Lockeia siliquaria* in the study area (all scale bars and squares are 10 mm). A) This base of bed (AM 7672) shows both *Lockeia* and *Ptychoplasma*-like elements, as well as a plug-shaped burrow. B) Total exploitation of the base of a sandstone bed by *Lockeia* traces. C) Enlargement from B showing the individual traces. D) One isolated trace from a siltstone bed top. E) Numerous hypichnial *Lockeia*, some of them serially aligned (cf. *Ptychoplasma*) covering the base of a sandstone bed (AM 7683). A, B and E are from the upper Weltevrede Formation, D is from the upper Rooirand Member of the Witpoort Formation.

Ichnogenus *Nereites* MACLEAY 1839*Nereites* isp.

Fig. 6.11 A–G

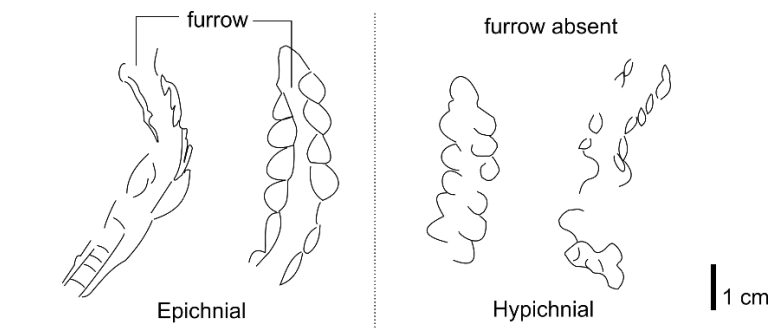


Figure 6.10: Line drawings of specimens from this study illustrating the morphology of epichnial and hypichnial *Nereites*, showing that a medial furrow occurs in the epichnial forms, which are otherwise similar in size and variation.

Occurrence– Restricted to one stratigraphic unit in the study area (Rabbit Ridge upper section), an upward-fining succession of thin-bedded, mud-draped sandstone with parallel, undulatory and flat lamination. *Nereites* occurs at the sandstone–mudstone interface in epirelief and hyporelief, in very high density to low density. Associated traces include ‘compound trace 1’. *Planolites* isp., *Palaeophycus tubularis* and indeterminate vertical burrows.

Horizon – Lower Rooirand Member

Material– **RR:** AM 7789 (7 hypichnial specimens); AM 7782 (several hypichnial and epichnial specimens), AM 7790 (three epichnial specimens), AM 7785 (two hypichnial specimens)

Description– Horizontal, irregularly wandering burrows comprising one, two, or more rows of round, oval or lanceolate lobes with or without a median furrow (furrow only seen in epichnial forms; Fig. 6.10). The trace is bedding parallel for short distances, usually

beginning and ending with a single lobe, indicating shifting the vertical position in the substrate (Seilacher 2007). Lobes are similar in composition to the matrix.

In the epichnial expression (top surfaces of sandstone beds), the trace has a median furrow that is bordered by lobate ridges (Fig. 6.11A and B). The furrow has a u-shaped profile, and it is either smooth, or contains a slight transverse ornamentation. The lobes contain oval to lanceolate pustules up to 8.5 mm in diameter that are arranged in an alternate pattern, with the long axis oblique to the furrow. Trace width is irregular, 7–18 mm wide, with a median furrow of 3.5–8.5 mm wide, comprising approximately half of the total width. Traces reach a maximum preserved length of 55 mm, meandering in an apparently non-systematic way, with acute turns, open curves and straight sections.

Hypichnial variants lack a central furrow but are formed of either a single row (uniserial) or two (or more) rows of pustules (biserial/multiserial) (Fig. 6.11C–G). The trace is highly variable in expression and these forms intergrade. Traces are 8–15.5 mm wide and up to 120 mm long with pustules up to 8.5 mm in diameter. The biserial form is similar to the epichnia described above, with alternating oval or lanceolate pustules. Uniserial forms have pustules that are elongated transverse to trail direction. Traces do not self-cross, and appear to avoid others of the same type (Fig. 6.11G).

Remarks– *Nereites* is a complex backfill structure characterised by a central tunnel (faecal string) with an envelope of reworked sediment lobes (Uchman 1998). *Nereites* is a vagile deposit feeding trace, considered to represent combined locomotion and feeding (paschichnia) of a worm-like animal (Chamberlain 1971; Seilacher 2007), however Rindsberg (1994) argued that worms, molluscs, arthropods or echinoderms are possible tracemakers for this ichnogenus. Although giving its name to the *Nereites* ichnofacies, typically representing deep-marine, abyssal settings (MacEachern *et al.* 2010), *Nereites* is eurybathic, and is common in shallow-marine and intertidal settings, particularly in the

Palaeozoic (Mángano *et al.* 2002b). Various *Nereites* species known from deep-marine deposits (Chamberlain 1971; Uchman 1998), are also recognised in shallow-marine and intertidal settings (Mángano *et al.* 2002b). Shallow-marine and intertidal forms tend to be irregular and unsystematic by comparison with the systematic, patterned meanderings of deep-marine flysch forms (Seilacher 1986; Rindsberg 1994; Mángano *et al.* 2002b).

According to Seilacher (1986, p. 68), the “aspect (of *Nereites*) changes in different modes of preservation.” Thus, ichnospecies of *Nereites* are commonly found to intergrade (Chamberlain 1971; Uchman 1995; Mángano *et al.* 2000) and this has contributed to multiple synonymies performed by various workers (e.g. Orr and Pickerill 1995; Uchman 1995). Epichnial versus hypichnial preservation is an important control on the morphology of *Nereites* which may record a spectrum of morphologies reflecting a single tracemaker (Chamberlain 1971; Seilacher 1986; Uchman 1995). This appears to have been the case in the studied sample where forms resembling several ichnospecies intergrade. Some of the material resembles *N. jacksoni* in having sub-circular lobes, whereas other examples have lanceolate lobes resembling those of *N. missouriensis* (Rindsberg 1994; Uchman 1998; Mángano *et al.* 2002b). Debates about *Nereites* taxonomy (reviewed by Mángano *et al.* 2000, 2002) are not entered here and the traces are not designated at the ichnospecies level.

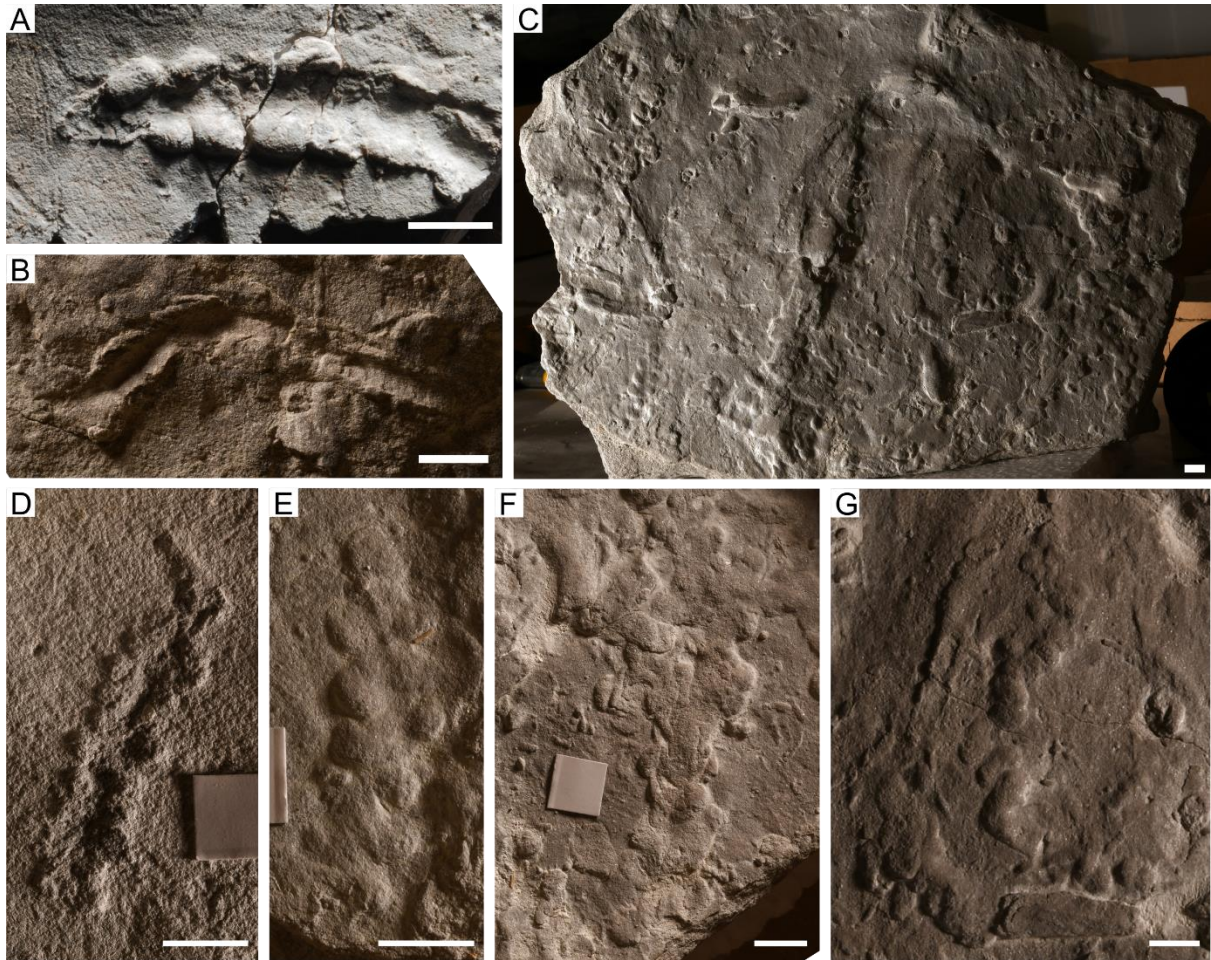


Figure 6.11: *Nereites* isp. in the lower Witpoort Formation (all scale bars are 10 mm). A–B) Epichnial forms, C–G) hypichnial forms. E, F and G are enlargements of traces from C.

***Ichnogenus Palaeophycus* HALL 1847**

Palaeophycus tubularis Hall 1847

Fig. 6.12 A–C

Material– HP: numerous specimens (not collected); **RR:** AM 7802 (one specimen); AM 7794 (a + b) (one specimen) ; AM 7782 (one specimen)

Horizon– Upper Weltevrede Formation, lower Rooirand Member (Witpoort Formation)

Occurrence– *Palaeophycus* is abundant at certain limited horizons in the upper Weltevrede Formation, occurring in mottled fine-grained sandstone and parallel laminated fine-grained sandstone. It occurs sporadically, in low abundance, in the lower Witpoort

Formation in silty carbonaceous sandstone, and in fine-grained sandstones in the Rabbit Ridge upper section.



Figure 6.12: *Palaeophycus tubularis* in the upper Weltevrede Formation (A–B) and in the lower Witpoort Formation (C) (all scale bars = 1 cm).

Description– Unbranched, bedding parallel or oblique, straight to slightly curved burrows with an elliptical cross section. Smooth walled, commonly with a thin (sub-millimetre) clay lining (Fig. 6.12A–C), or with a thick sandy lining (Fig. 6.13B). Fill is structureless and similar in texture to the matrix. Diameter is variable, usually 5–10 mm (up to ~20 mm).

Preserved as hypichnia and endichnia.

Remarks– *Palaeophycus* is a simple horizontal burrow with a lining (Pemberton and Frey 1982; Mángano *et al.* 2002b). *Palaeophycus tubularis*, the type ichnospecies, is distinguished by its smooth walls (Mángano *et al.* 2002b). It is separated from superficially similar *Planolites* by the presence of a lining and structureless fill that is identical to the host rock,

which is considered to indicate an open burrow that has been passively filled (Pemberton and Frey 1982). One of the most widespread ichnogenera reported in the literature, this is a facies-crossing ichnotaxon that ranges from the Precambrian to Recent (Gingras *et al.* 1999; Mángano *et al.* 2002b). It is thought to represent an open, permanent domicile such as, for example, predatory polychaetes inhabit (Pemberton and Frey 1982).

Ichnogenus *Phycodes* Richter 1850

Phycodes cf. *palmatus* Hall 1852

Fig. 13A–C

Material– WF: AM 7606 a, b (four specimens), AM 7610 (one specimen – burrow and chamber), **RR:** AM 7782 (one specimen)

Horizon– Lower Rooirand Member (Rabbit Ridge upper section) and upper Rooirand Member (Waterloo Farm)

Occurrence– This ichnospecies description is based on uncommon examples occurring on two rock slabs in the studied sample set; one from the Rabbit Ridge upper section and one from Waterloo Farm. Both occur in thin-bedded pinch and swell sandstone with muddy partings that underlie mudstone/shale. The Waterloo Farm specimens occur on a bedding surface that is covered with diminutive traces tentatively attributed to ‘compound trace 1’, from a similar stratigraphic horizon to the *Cruziana problematica* described above. The Rabbit Ridge example is associated with *Nereites* isp., *Scalichnus lepusaures*, and *Palaeophycus tubularis*.

Description– Horizontal, branching burrow system comprising smooth or irregularly walled hypichnial ridges that are highly variable in morphology. Branching is palmate, with up to five lobate, finger-like projections ramifying from a proximal tunnel. Branches are 7–13 mm in diameter, emerge at an oblique or perpendicular angle to the master tunnel, are either

recurved or straight with round terminations. Recorded burrow system lengths are between 78–135 mm. The trace may also terminate in an enlarged chamber, with one example on AM 7606 being 25 mm in diameter (Fig. 6.13C). Burrow fill is similar in composition to the overlying bedding surface.

Remarks– *Phycodes* is reserved for bundled sets of burrows emanating from a proximal master tunnel (Han and Pickerill 1994; Uchman 1998). Twelve ichnospecies are recognised, incorporating a wide variety of morphologies (Han and Pickerill 1994). *Phycodes* is considered to represent deposit feeding by a vagile animal, possibly a worm or an arthropod (Han and Pickerill 1994). Annelids have been suggested as likely producers (Seilacher 1955; Mángano *et al.* 2002b). *Phycodes* is most common within the shallow-marine *Cruziana* ichnofacies, but also occurs in non-marine, marginal-marine and deep-marine settings (Mángano *et al.* 2002b). The low number of branches and palmate form tentatively suggest assignment to *P. palmatus* (Fillion and Pickerill 1990; Mángano *et al.* 2002b), however, the traces could also represent incomplete *P. templus*, which presents a similar style of branching, but is a more horizontally extensive burrow system (Han and Pickerill 1994).

Ichnogenus *Planolites* NICHOLSON 1873

Planolites *ispp.*

Fig. 6.14 A–G

Material– **HP:** AM 7647 (one specimen), AM 7645 (one specimen); **RR:** R2 087 (one specimen), R2 089 (one specimen); uncollected material from various stratigraphic horizons

Horizon– Occurs throughout the studied stratigraphic interval (upper Weltevrede Formation to Skitterykloof Member) but is not recognised in the Perdepoort Member

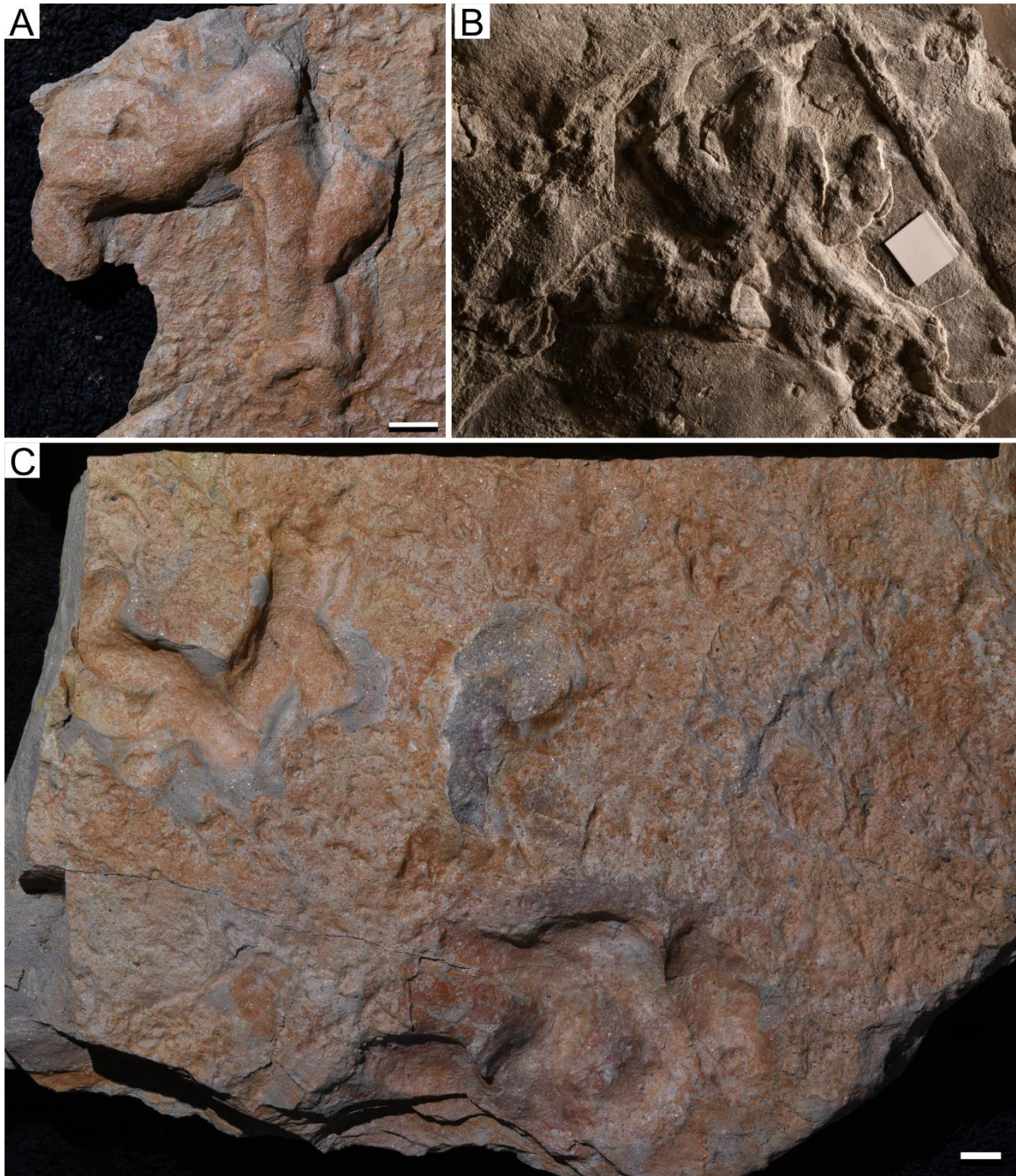


Figure 6.13: *Phycodes* cf. *palmatus* from the Witpoort Formation (all scale bars are 10 mm). A, C) Specimens from Waterloo Farm (AM 7606), A represents the counterpart to the specimen at top left of photograph C. B) Specimen from Rabbit Ridge (AM 7782). *Palaeophycus tubularis* is seen at the top right of the photograph.

Description– Encompasses a range of simple horizontal burrows lacking a lining.

Planolites is common in muddy facies at various horizons, but almost always as indistinct, small-sized (usually less than 5 mm in diameter), short, poorly preserved structures that are seldom collected (Fig. 6.14A, B). Large hypichnial forms that are welded to the overlying sandstone bed occur at a sandstone–mudstone contact in the lower Witpoort Formation at Howison’s Poort. The burrows there intergrade with *Ptychoplasma* (Fig 6.14C). Near the top of the Witpoort Formation, monospecific *Planolites* occur in great abundance in gravelly sandstone facies (Fig. 6.14E, F). They are parallel-sided irregularly meandering burrows 10–15 mm wide and up to 8 cm long.

Remarks– This ichnotaxon encompasses unbranched, unlined horizontal burrows (Keighley and Pickerill 1995). Pemberton and Frey (1982) defined *Planolites* as a backfilled structure that differs in texture and composition from the hostrock, indicating a mobile deposit feeding burrow. This approach was, however, countered by Keighley and Pickerill (1995) in that actively filled burrows do not necessarily differ from the hostrock in texture or composition, and, in converse, a distinct burrow fill does not exclude passive filling. The diagnostic application of this ichnogenus is limited in that it probably encompasses a variety of tracemaking organisms, it occurs throughout the Phanerozoic, and it occurs in almost every sedimentary environment (Buatois *et al.* 2016). Straight forms of *Planolites* may, furthermore, be difficult to separate from tool marks (Fig. 6.14G). Intergradation with *Ptychoplasma* indicates that bivalves may have been responsible for some *Planolites* in the study area.

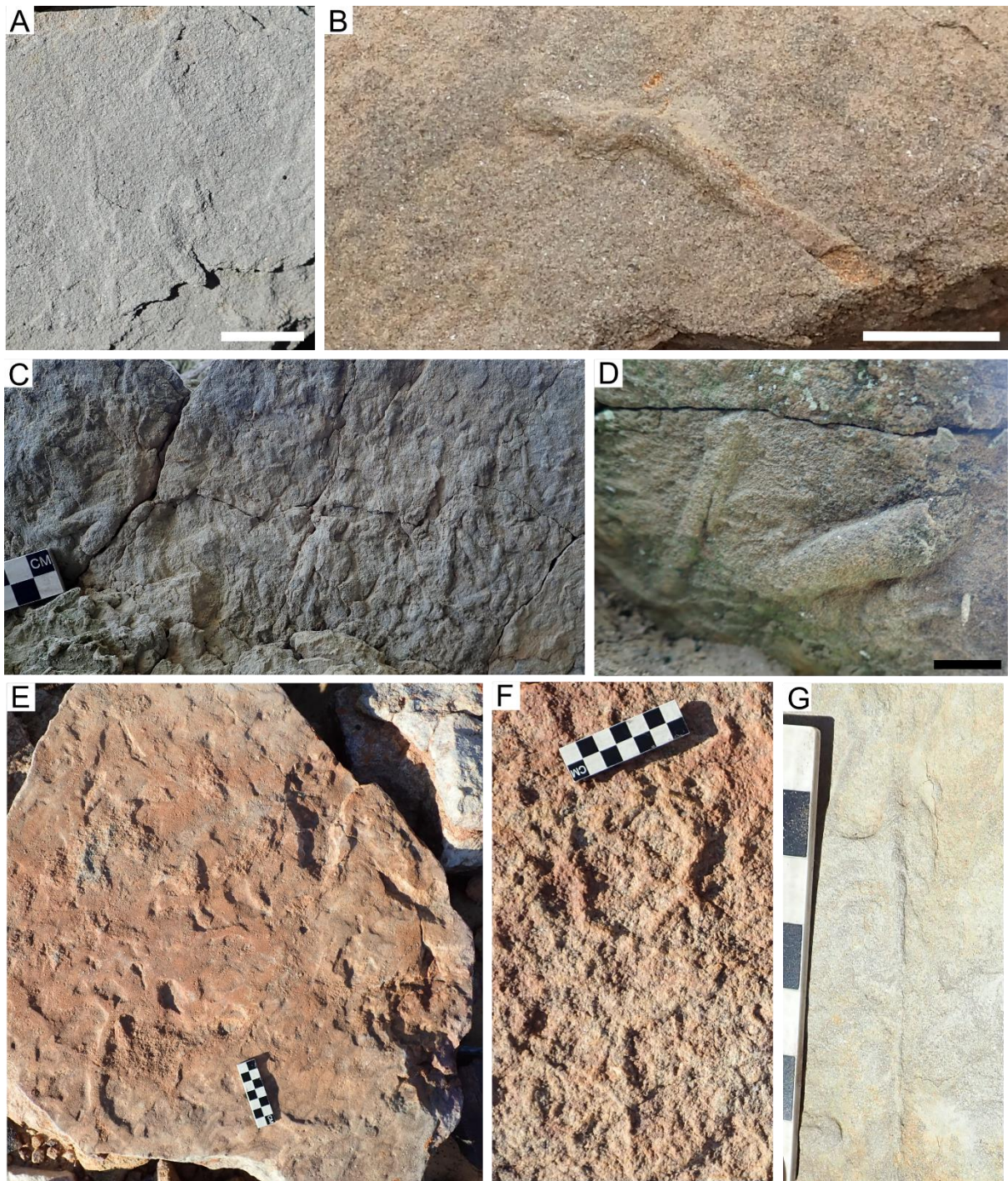


Figure 6.14: *Planolites* from the Witpoort Formation photographed at outcrop (all scale bars and squares are 1 cm). A–B) Small examples from siltstone and fine grained sandstone facies. C–D) Irregular hypichnia, with examples to the left of the photograph in C intergrading with *Ptychoplasma*. E–F) Monospecific *Planolites* beds in the Skitterykloof Member. G) *Planolites*-like structure that is parallel to the palaeoflow direction, which might be a tool mark.

Ichnogenus *Psammichnites* TORELL 1870

Psammichnites implexus (Rindsberg 1994)

Fig. 6.15 A–C

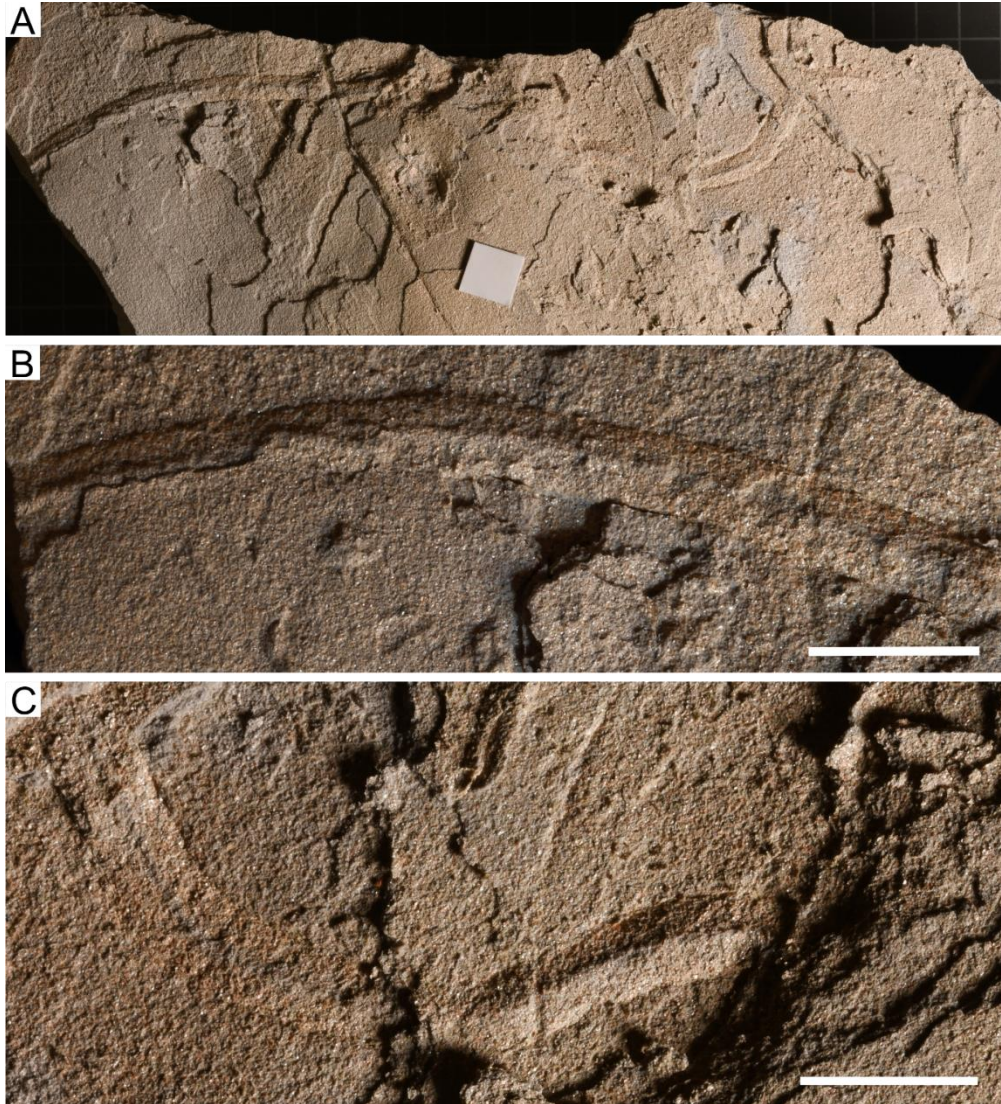


Figure 6.15: *Psammichnites implexus* from the lower Witpoort Formation Rabbit Ridge locality (all scale bars = 1 cm) A) general view of the slab (AM 7844). B–C) close up of the specimens, note the slightly darker colour of the median ridge.

Material – RR: AM 7844 (two specimens)

Occurrence – *Psammichnites implexus* is described based on specimens from one slab of parallel-laminated fine-grained sandstone interbedded with mudstone at Rabbit Ridge. It is associated with walled vertical burrows (*Lingulichnus verticalis*) and is cross-cut by diminutive horizontal burrows.

Description – A gently meandering, bedding-parallel epirelief, very consistent in width, with a smooth bifacial upper surface culminating in a sharp, continuous medial ridge ~1 mm wide. The ridge is composed of slightly darker sediment than the matrix, whereas the remainder of the trace appears identical to the matrix in composition. Where the trace disappears beneath an overlying bedding plane, a faint trace of this ridge cuts the overlying horizon. Traces are 4.5–5.2 mm wide.

Remarks– *Psammichnites* represents the infaunal activities of a vagile animal that maintains connection with the surface using a dorsally-orientated siphon (Mángano *et al.* 2002a; Seilacher 2007). Pazos and Gutierrez (2023) argued that all post-Cambrian *Psammichnites* belong to *Olivellites* and should be restricted to the type species, however, *Olivellites* has also been considered a junior synonym of *Psammichnites* (Mángano *et al.* 2002a; Mángano *et al.* 2022). The present traces are similar to *Psammichnites implexis*, with the apparently smooth, two-faced upper surface culminating in a continuous dorsal ridge, and the small diameter (Rindsberg 1994; Mángano *et al.* 2002a; Morell and Denayer 2019). The two specimens here studied lack the characteristic scribbling tendency of *P. implexus*, but this does not exclude them from the ichnospecies as non-scribbling examples are known (Mángano *et al.* 2002a). *Psammichnites implexis* occurs in intertidal and brackish water settings in the Devonian and Carboniferous Periods (Rindsberg 1994; Mángano *et al.* 2002a, 2002b; Alonso-Muruaga *et al.* 2013; Morelle and Denayer 2020). A siphonate mollusc is indicated as a likely tracemaker for this ichnospecies (Mángano *et al.* 2002a).

Ichnogenus *Ptychoplasma* Fenton and Fenton 1937

Ptychoplasma vagans (Ksiazkiewicz, 1977)

Fig. 6.16 A–F

Material– **CH:** AM 7746 (one specimen), AM 7747 (multiple overlapping specimens), AM 7744 (a, b) (one specimen), AM 7748a, AM 7748b (multiple overlapping specimens), AM 7749 aii (one specimen); **HP:** AM 7664 (a, b), AM 7647 (two specimens), AM 7672 (two specimens).

Horizon– Upper Weltevrede Formation to uppermost Rooirand Member

Occurrence– *Ptychoplasma* is common in the upper Weltevrede Formation and in limited, discreet horizons in the lower–middle Witpoort Formation. It forms dense, sometimes monospecific accumulations at the bases of FA 2c heterolithic sandstone beds and occurs in lower abundance in argillaceous rocks. At Coombs Hill the trace is associated with abundant sub-vertical tubes and articulated bivalve shells in the sandstone strata underlying mudstone, as well as within mudstone itself. Co-occurs with *Planolites*, *Archaeonassa*, *Scalichnus*, indeterminate vertical burrows and unspecified diminutive traces.

Description – Sub-horizontal series of amygdaloid hypichnial ridges or, less commonly, epichnial grooves. Ridges are externally smooth, occasionally preserving a slight chevronate ornament, subsiding at both ends, with variable degrees of overlap from isolated to contacting. In places the ridges are amalgamated to a degree that they are indistinguishable, and resemble *Planolites* (Figs 6.14C, 6.16C). Individual ridges are 8–23 mm long, between a third and half as wide as long, up to 4 mm deep. In cross section, ridges are carinate to semicircular and may be connected to vertical or oblique tubes extending into the overlying sediment (Fig. 6.16E). The trace is composed of identical material to the overlying bed, to which the amygdalar bodies are welded.

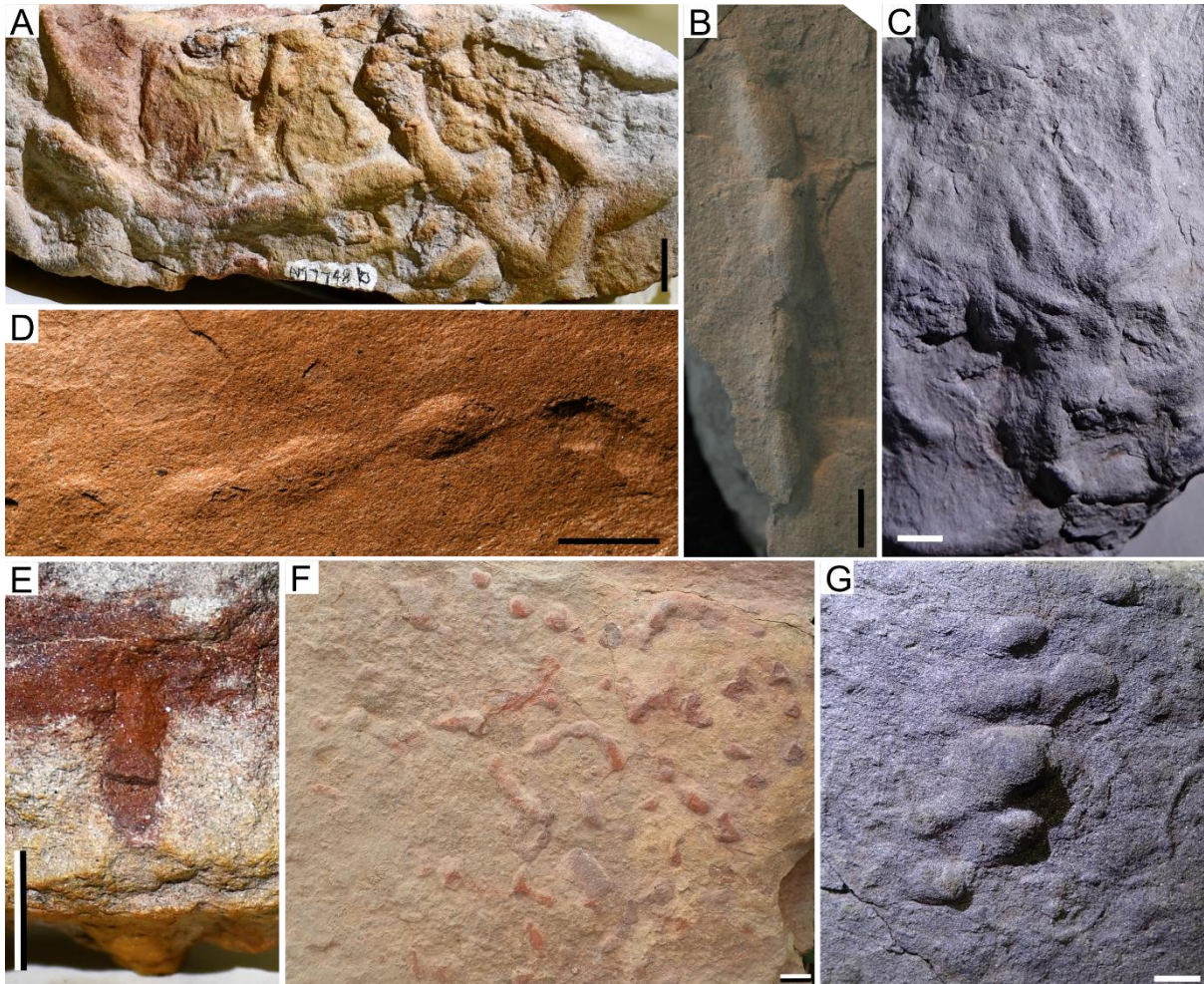


Figure 6.16: *Ptychoplasma* from the upper Weltevrede Fm and Witpoort Fm (scale bars = 10 mm).
A) AM 7748b. B) AM 7744b. C) AM 7747. D) A 7647. E) Vertical profile showing a cylindrical tube overlying *Ptychoplasma* (AM 7748a). F) *P. vagans* intergrading with *P. conica* (uncatalogued slab from Seven Fountains). G) *P. conica* from Waterloo Farm (AM 18038)

Ptychoplasma conica Pienkowski and Uchman 2009

Fig. 6.16 G, F

Material– SF: uncatalogued slab (at least four specimens). **WF:** AM 18038 (one specimen)

Occurrence– Co-occurs and intergrades with *P. vagans*.

Horizon– Upper to uppermost Rooirand Member

Description– Similar to *P. vagans* but differs in that the hypichnial ridges are round rather than amygdaloid. Ridges are less than 10 mm in diameter.

Remarks– The serially arranged, almond-shaped or oval ridges are consistent with the ichnogenus *Ptychoplasma* (Uchman *et al.* 2011). Individual segments resemble *Lockeia*, and were thus designated as *Lockeia serialis* by Seilacher and Seilacher (1994), which was, however, shown to be a *nomen nudum* (Pieńkowski and Uchman 2009). According to Uchman *et al.* (2011) *Lockeia* is reserved for resting traces, whereas *Ptychoplasma* is a locomotory structure. Some accepted *Lockeia* ichnospecies such as *L. cunctator* (Schlirf *et al.* 2001) and *L. ornata* (Mángano *et al.* 2002b) show clear evidence of intermittent locomotion. Three ichnospecies of *Ptychoplasma* are recognised, which are distinguished by the shape and continuity of the segments. The present traces are attributed to two ichnospecies. *Ptychoplasma vagans* is distinguished by discontinuous amygdaloid segments (Uchman *et al.* 2011). This ichnospecies intergrades with more overlapping forms that could be referred to *P. excelsum* on this basis. *P. conica* is distinguished by its non-overlapping conical ridges that occur in serial alignment (Pieńkowski and Uchman 2009). Wedge-foot bivalves are likely producers (Uchman *et al.* 2011; López Cabrera *et al.* 2019), though cleft-foot nuculoid bivalves are also potential tracemakers (Mángano *et al.* 1998). *Ptychoplasma* occurs with isolated *Lockeia* in the present study area, which resemble the individual *Ptychoplasma* segments, indicating a common tracemaker and mechanism of action for the locomotion and resting traces. One of the undescribed bivalve forms from Coombs Hill, the ‘hard-shelled’ morphotype (Harris *et al.* 2021a), occurs in abundance with *Ptychoplasma vagans* and is a plausible tracemaker. *Ptychoplasma excelsum* is a shallow to marginal-marine ichnospecies, widespread in intertidal facies in the Carboniferous of North America (Uchman *et al.* 2011). Wedge-foot bivalves are likely producers (Uchman *et al.* 2011; López Cabrera *et al.* 2019), however, cleft-foot nuculoid bivalves are also potential tracemakers (Mángano *et al.* 1998).

Ichnogenus *Rusophycus* Hall 1852

Rusophycus carbonarius Dawson 1864

Fig. 6.17 A–C

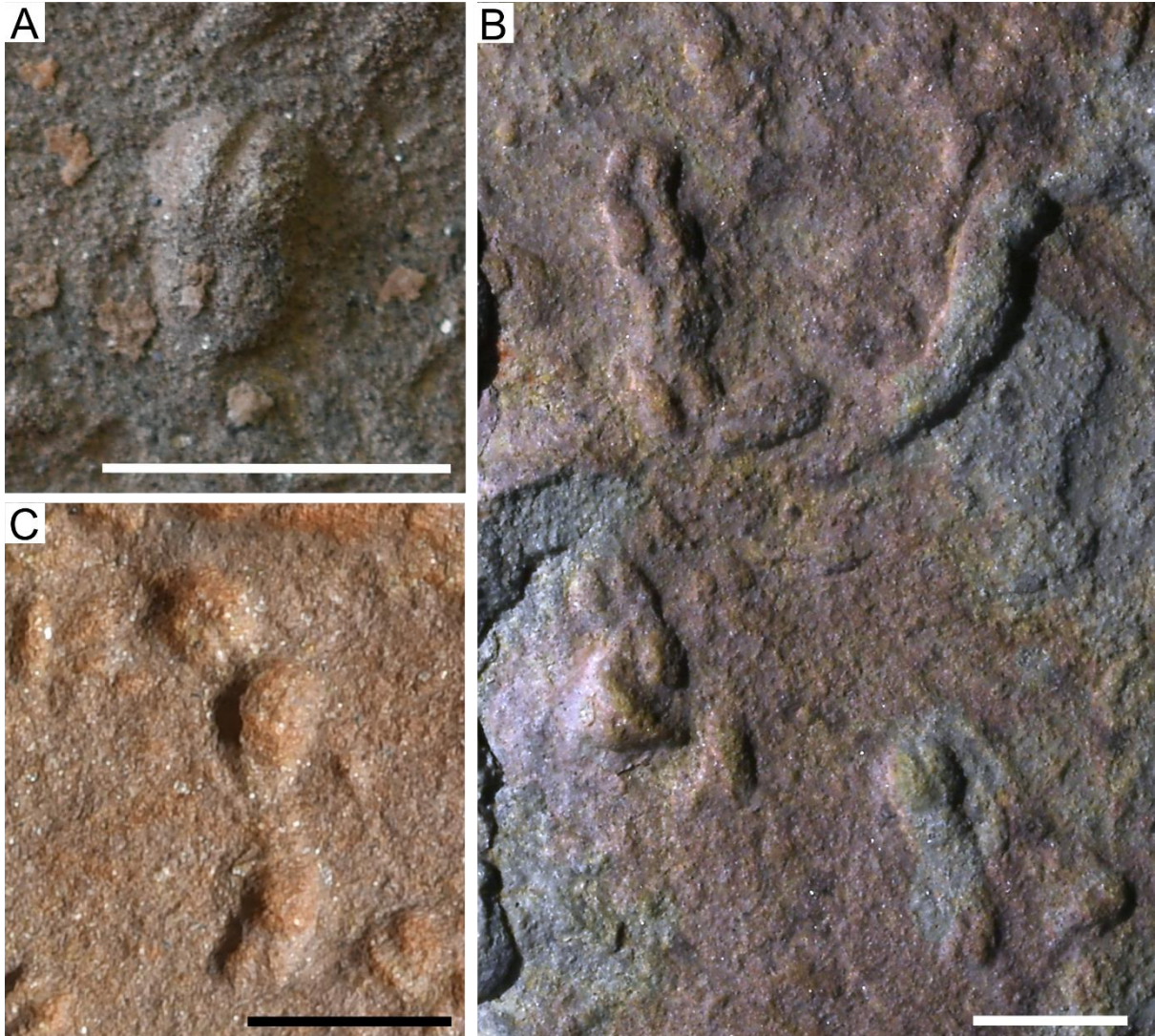


Figure 6.17: *Rusophycus carbonarius* from the Waterloo Farm locality in the Witpoort Formation (all scale bars = 10 mm). A) Two superimposed resting structures, note the faint transverse scratch marks on the upper trace in the photograph (AM 7625). B) *Rusophycus* with various degrees of overlap, with superimposed examples producing *Cruziana*-like structures. (AM 5402). C) Overlapping *Rusophycos* forming part of a linear series (AM 7609).

Material– WF: AM 7625 (10's of specimens), AM 7610 (6 specimens), AM 7609 (10's of specimens), AM 5402 (c, e, f, j, l, n, v, w, y, z) (10's of specimens), AM 5403 (10's of specimens)

Occurrence– Ubiquitously occurring with *Cruziana problematica* within thin-bedded quartz wacke underlying the MFL at Waterloo Farm (see above). It is the dominant ichnogenus, which co-occurs with ‘compound trace 1’ and small *Lockeia*.

Description– Coffee-bean shaped, bilobed hypichnial mounds (rarely epichnial grooves), 3–6 mm wide, up to 6 mm long, comprising a pair of parallel ridges sharply separated by a shallow, straight median furrow ≤ 1 mm wide. A series may be superimposed forming a *Cruziana*-like trace. The lobes are smooth or have a faint transverse ornamentation (Fig. 6.17A). The traces occur in sub-parallel alignment on all studied slabs, isolated or superimposed in linear series (Fig. 6.17B, C).

Remarks– *Rusophycus* and *Cruziana* exhibit a gradational relationship and almost ubiquitous co-occurrence in the rock record, hence, their synonymy has been proposed (Seilacher 1970), however, most researchers maintain them as separate (e.g. Keighley and Pickerill 1996; Mángano *et al.* 2002b). *Rusophycus* is separated from *Cruziana* by having a length to width ratio of less than 2:1 (Keighley and Pickerill 1996). The trace is interpreted as the short-lived resting trace of a vagile arthropod (Trewin 1976; Pollard 1985; Kesidis *et al.* 2018). *Rusophycus* is common in continental, intertidal, and shallow-marine settings (Mángano *et al.* 2002b), and is recorded in deep-marine, pre-Silurian strata (Pickerill 1995). Although commonly ascribed to trilobites, non-marine forms, evidently not produced by trilobites, are indistinguishable (Seilacher 1970; Bromley and Asgaard 1979; Keighley and Pickerill 1996). *Rusophycus* taxonomy is largely based on the morphology of the scratch marks (Keighley and Pickerill 1996). The small size and parallel lobes that are smooth or with poorly developed transverse ornament are typical for *R. carbonarius* (Keighley and Pickerill 1996; Schlirf 2005). *R. eutendorfensis* differs in having radial, typically bifid scratch marks that extend beyond the margins of the lobes (Keighley and Pickerill 1996). *R. stromnessi* is characterised by ‘mamillate’ lobe endings, which excludes the Waterloo Farm

material from the species (Trewin 1976; Keighley and Pickerill 1996). *Cruziana problematica* is associated with *R. carbonarius* in marginal-marine settings, such as in the Late Devonian of New York State (Miller 1979), in the Early Devonian of Antarctica (Bradshaw 2010) and in estuarine deposits of Late Cambrian age in South America (Mángano *et al.* 2023), but is also found in non-marine settings, such as in the Carboniferous of eastern Canada (Keighley and Pickerill 1997). Likely producers of small, non-marine or intertidal *Rusophycus* include isopods and amphipods (Bromley and Asgaard 1979; Pollard 1985; Schlirf *et al.* 2001; Mángano *et al.* 2002b).

Ichnogenus *Selenichnites* (*Selenichnus*) Romano and Whyte 1987

Selenichnites cf. *rossendalensis* Hardy 1970

Fig. 6.18, 6.19 A–C

Material– RR: AM 7844 (approx. 50 traces on one slab)

Horizon– Lower Rooirand Member

Occurrence– Lower bounding surface of an erosively-based channel sandstone unit (FA 2a) overlying siltstone. Around 50 traces are recorded on a roughly 90 cm x 30 cm bedding plane. All traces are orientated similarly in a south-westerly (offshore) direction, parallel to the palaeocurrent direction based on ripple-toe marks on the slab (Figs 6.18, 6.19C). A few tiny (<1 mm wide) bedding-parallel, horizontal burrows wander across the surface, and a vertical cylindrical structure intersects the bedding plane. Equilibrium traces occur in an immediately underlying fine-grained, thin sandstone interbed.

Description– Smooth, bilaterally symmetrical, crescent-shaped hypichnial mound, comprising two reniform lobes that are in contact anteriorly, sometimes joined anteriorly forming a horseshoe-shaped structure, 10–12 mm wide, slightly wider than long. Some of the traces have an acuminate median posterior ridge, up to 8 mm long, emerging from between

the lobes where 1–2 mm wide. The texture is similar to the host rock. Seven examples on the slab show the traces in linear alignment, with up to four traces in a series (Fig. 6.19A).



Figure 6.18: Three serially aligned *Selenichnites* travelling downcurrent. The traces are brief resting intrusions into the sediment by a small limulid. Very fine backfilled burrows pass across this surface, and a vertical cylindrical structure truncates the current ripple markings.

Remarks– Occurrence of bilaterally symmetrical traces in serial alignment indicates a combination of resting and locomotory behaviour in a motile animal. That all traces are orientated similarly suggests several such animals moving in unison, forming shallow burrows in the sediment for resting purposes. Only the resting traces are preserved, and it is assumed that the animals were locomoting at a slightly higher bedding horizon, their locomotory trails having not preserved.

The horseshoe shape, dimensions, posterior ridge and serial alignment identify the traces as *Selenichnites* and are separated from morphologically similar ichnogenera

Limulicubichnus Miller 1982, which is a teardrop-shaped, epichnial depression tapering to a

posterior groove and *Crescentichnus* Romano and White 2015, which lacks the characteristic posterior ridge or groove (Romano and Whyte 2015). Among *Selenichnites* species, only two are known to occur in serial alignment, *S. hundalensis* (the type ichnospecies) and *S. rossendalensis*. The Rabbit Ridge traces are distinguished from *S. hundalensis* in lacking a distinct subtriangular segment anteriorly. They are almost identical to *S. rossendalensis*, but lack the posterior subparallel rows of *en eschelon* markings that form part of a revised diagnosis for the ichnospecies (Romano and Whyte 2015). The species diagnosis was again emended to include specimens with or without scratch marks and posterior ridge (King *et al.* 2019). This extends a long and complicated taxonomic history: originally *Kouphichnium rossendalensis*, the ichnospecies was transferred to *Limulicubichnus* by Miller (1982) and then to *Selenichnus* Romano and Whyte 1978, which became *Selenichnites* (Romano and Whyte 2015).

Selenichnites is interpreted as the resting traces of limulids (horseshoe crabs) (Romano and Whyte 2015; King *et al.* 2019), but any arthropod with a semilunate cephalon, such as for example, trilobites or eurypterids, can potentially form *Selenichnites* (Gibb *et al.* 2011; King *et al.* 2019). The elongate posterior ridges in the present examples strongly support limulids as tracemakers (Buhler and Gray 2017; Romano 2021).

S. rossendalensis is a common ichnospecies in ancient marginal-marine deposits; it is reported from Carboniferous inner estuary deposits in Alabama (King *et al.* 2019) and in Permian tidal-flat deposits (Lerner and Lucas 2015). Some occurrences in strata previously interpreted as non-marine are now thought to have been marine influenced, such as those from the Middle Jurassic of Yorkshire (Romano 2021) and the Pennsylvanian Joggins Formation in Nova Scotia (Rygel *et al.* 2014). The ichnospecies has been considered as a shoreline marker in otherwise non-marine Palaeozoic deposits (Eagar *et al.* 1985; Lerner and Lucas 2015).

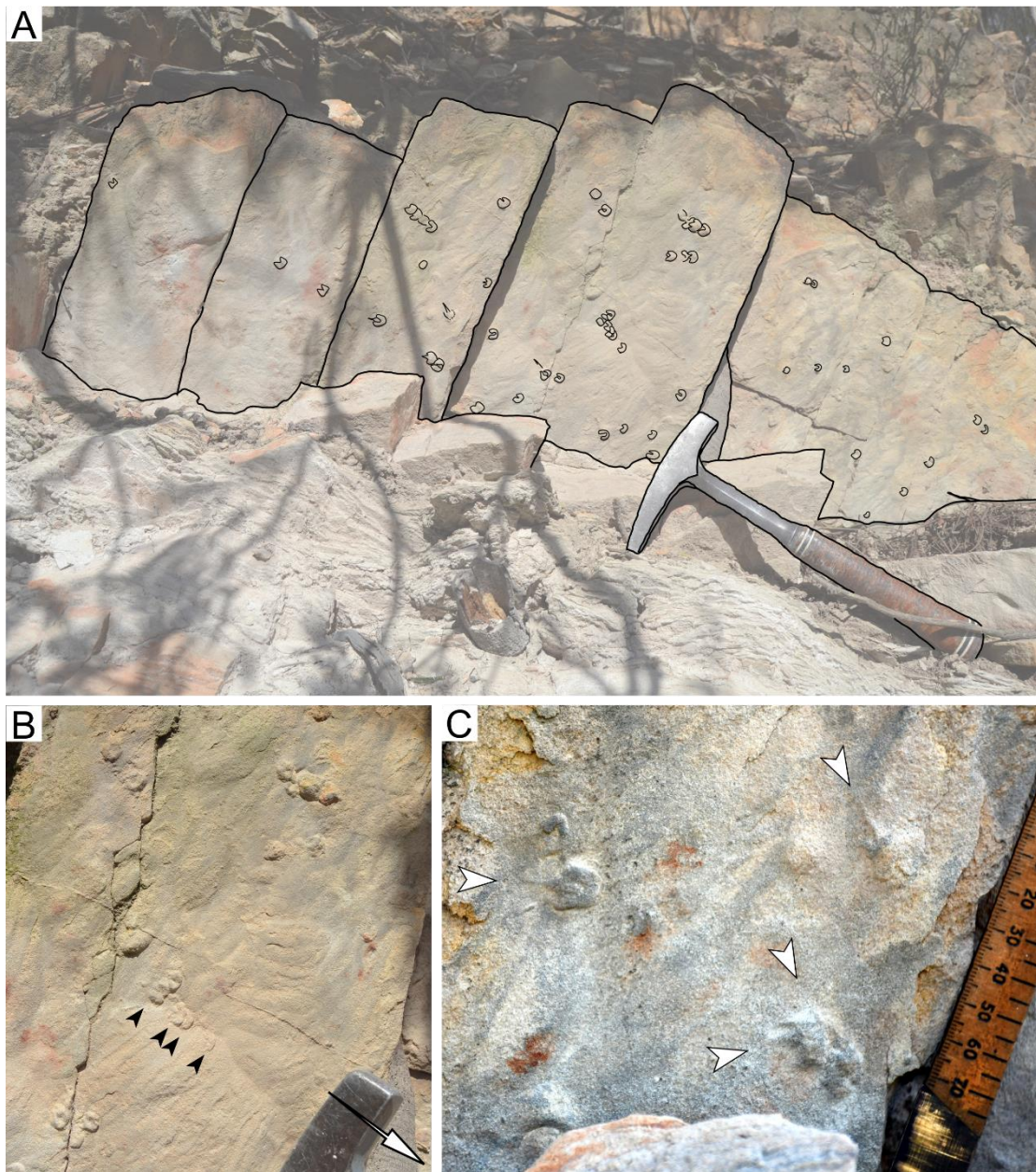


Figure 6.19: *Selenichnites rossendalensis* from the lower Witpoort Formation at Rabbit Ridge. A) Photograph and line drawing of the slab and traces (base of bed). B) Enlargement with black arrows indicating four traces in serial alignment. White arrow indicates the inferred palaeocurrent direction, noting that the rib and furrow marks in base of bed view formed in the convex direction. C) Enlargement with white arrows indicating traces that have a median posterior ridge (scale increments in cm).

Ichnogenus *Scalichnus* HANKEN, BROMLEY AND THOMPSON 2001

Scalichnus phiale Hanken, Bromley and Thompson 2001

Fig. 6.20 A, B

Material– CH: AM 7741aii (two specimens in profile), AM 7749ai (two specimens in profile and two specimens in plan), AM 7749aii (two specimens in profile)

Occurrence– Occurs in laminated sandstone with mud-drapes (facies Sc) with horizontal and low angle, rhythmic, cross-laminae. *Archaeonassa* cf. *fossulata* and vertical burrows including *Scalichnus lepusaures* and ‘plug-shaped burrows’ (equilibrichnia) also occur at this horizon, a metre-thick heterolithic interval underlying mudstone (Inner Outcrop horizon) at Coombs Hill.

Description– Bottle-shaped structure with a lined sub-vertical tube underlain by a plug-shaped (usually retrusive) structure with upward-concave laminae. Due to their sub-vertical dispositions, profile views only show this tube partially in the studied sample. One example is 4 mm wide and is visible for a length of 12 mm (Fig. 6.20B). It is inferred to connect with a tube intersecting the bedding surface 3 cm above its base. The retrusive structure is 1–2 cm in diameter and is highly variable in expression. On associated bedding planes, actively filled circular or sub-circular structures up to 3 cm in diameter represent the plan view of the basal structure.

Scalichnus lepusaures Zonneveld and Gingras 2013

Fig. 6.20 C–H

Material– CH: (plan view unless otherwise stated); AM 7740 (eight specimens), AM 7748 ai (four specimens) AM 7749 f (three specimens, one in profile), AM 7749 d (five specimens), AM 7749 ai (two specimens); **RR:** AM 7792, AM 7789 (at least six specimens); **WP:** Skitterykloof Member (tens of specimens observed at outcrop)

Horizon– Upper Weltevrede Formation at Waaipoort; lower Rooirand Member (Rabbit Ridge); Skitterykloof Member (Waaipoort)

Occurrence– Endichnia in sandstone beds, occupying heterolithic units. Abundant underlying Inner Outcrop, and in the uppermost strata of the Witpoort Formation at Waaipoort (WP 2 and WP 3): in sandy interbeds within a metre-thick, intensely bioturbated siltstone bed, and in the sandstone immediately overlying (WP 2). Similar traces are also recorded from the upper section at Rabbit Ridge. The trace is mostly recorded as plan-view examples intersecting bedding planes (Fig. 6.20D, G and H), and from rare profile views (Fig. 6.20C, E and F).

Description– Vertical or sub-vertical trace fossil comprising one, or a pair of, narrow cylindrical tubes (2–5 mm diameter) each enveloped by a mantle of reworked material (6-15 mm in diameter). Where in contact, the two mantled tubes are expressed plan view within a figure-of-eight shaped structure, a morphology that is widely repeated in the study area.

In the Coombs Hill material the pairs of tubes form depressions on sandstone bed tops, whereas in the Rabbit Ridge material the tubes are apparently deflected upwards, standing in positive relief on bed tops. One of them exposed in profile view is a downwards tapering, funnel shaped structure of oxidised sediment that interrupts the lamination of the host rock (Fig. 6.20E).

Remarks– *Scalichnus* was absorbed into *Siphonichnus* by Zonneveld and Gingras (2013), however, Knaust (2015) retained *Scalichnus* for the type ichnospecies, *S. phiale*, and included two other ichnospecies previously attributed to *Siphonichnus* (*lepusaures* and, tentatively, *sursumdeorsum*) in the ichnogenus. *Scalichnus* is therefore considered valid and distinct from *Siphonichnus* (Hanken *et al.* 2001; Knaust 2015; Buatois *et al.* 2017). *Scalichnus* is a sack- or bottle-shaped structure with concave-up equilibrium structures at the base, overlain by one or two tubes.

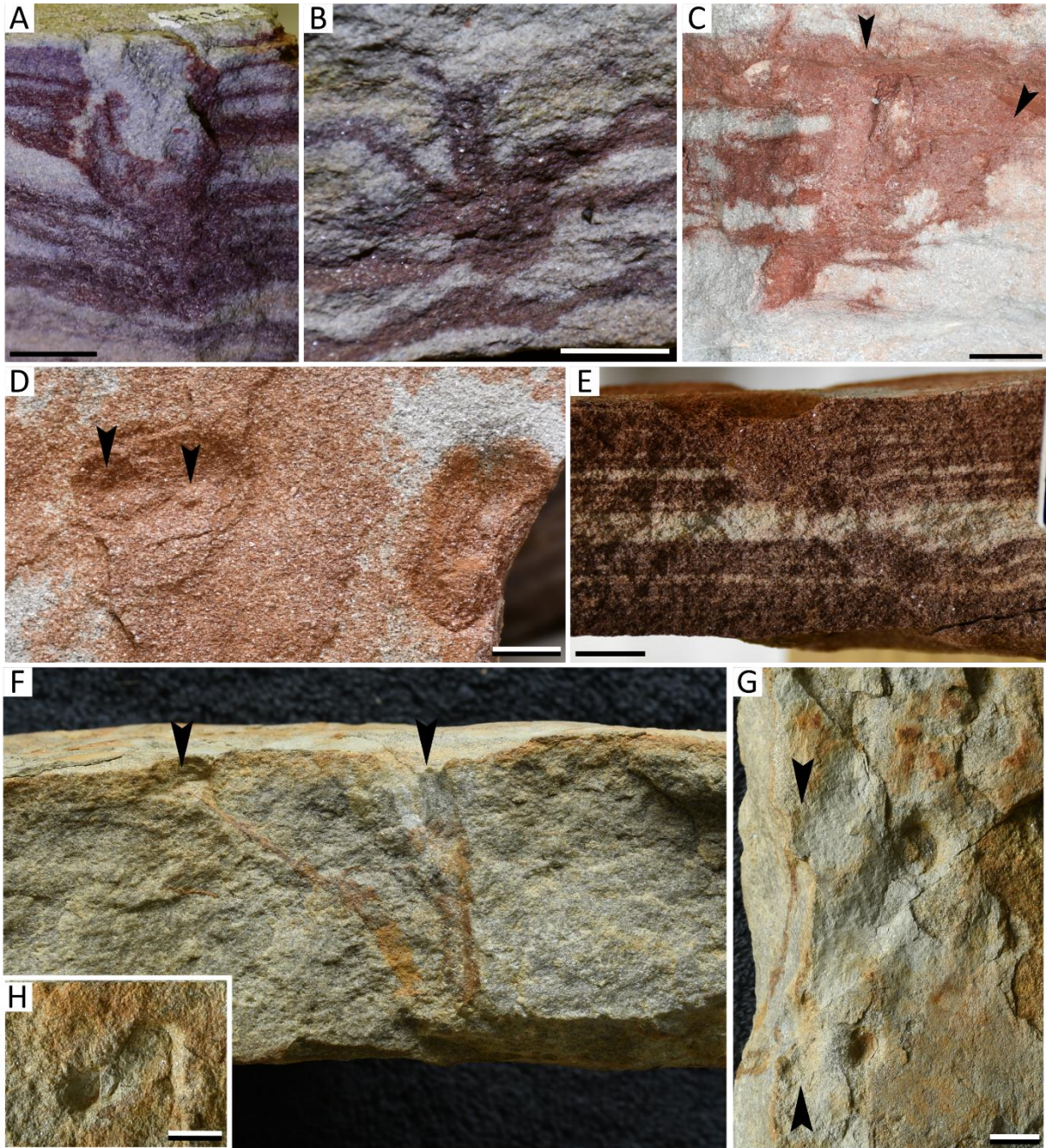


Figure 6.20: *Scalichnus* from the Witpoort Formation at Coombs Hill (A–E) and Weltevrede Formation (F–H) (all scale bars = 1 cm; arrows in C, D, F and G indicate paired tubes). A) Bottle-shaped retrusive structure with a single opening above (*S. phiale*), where it intersects a bedding surface (AM 7749aii). B) A single tube emerging from a plug-shaped furrow (AM 7749ai). C) Paired tubes above a basal ovalular structure (*S. lepusaures*) (AM 7748j). D) Paired tubes on a bedding surface forming a figure-of-eight shaped trace (AM 7749f). E) Profile view of the specimen broken through on the right of the slab in C. F–H) Slab from an unspecified horizon in the Weltevrede Formation (photographs courtesy of R. Nel).

Scalichnus is separated from *Siphonichnus*, firstly in the lack of a meniscate mantle bordering the central passively-filled tube, it rather having a wall-parallel lining, and secondly, *Scalichnus* is vertically-disposed, in contrast to the variable attitude of *Siphonichnus* (Knaust 2015). The type ichnospecies, *S. phiale*, is formed by the equilibrichnial movements of a sessile bivalve that maintains connection to the sediment surface via a vertically-orientated siphon (Hanken *et al.* 2001). The trace is highly variable in expression within the type and associated material from the Pliocene of Rhodes Island, Greece. Burrow fill is complex, resulting from repeated ‘yoyo’ movements and lateral wanderings. *Scalichnus lepusaures* differs from the type ichnospecies in that it has a pair of tubes rather than a single tube overlying the equilibrium structure (Knaust 2015).

Scalichnus lepusaures is hitherto unknown in its bedding plane expression. Mayoral *et al.* (2018) diagnosed a new ichnogenus and species from Pleistocene strata in Greece, *Alaichnus kabuverdiensis*, associated with fossils of the bivalve *Panopea*, the tracemaker for *S. phiale*, which they suggested could represent a bedding plane expression of a *S. lepusaures*-like trace with combined horizontal and protrusive spreite. In this study the paired tubes do not form spreite, but occur in conjoined or separated pairs. They bear a resemblance to pairs of tubes that extend upwards from the compound trace fossil *Hillichnus lobosensis*, representing the markings of a tellenid bivalve’s inhalant and exhalant siphons (Bromley *et al.* 2003). It is predicted that the presently described material will be found to emerge from an underlying equilibrium and/or locomotory structure.

Ichnogenus *Undichna* ANDERSON 1976*Undichna* ispp.

Fig. 6.21 A–C

Occurrence– *Undichna* ispp. are common in black shale facies from Waterloo Farm MFL. They occur on slightly coarser-grained, silty surfaces that occur at regular (centimetre-scale) intervals in the unit. Synaeresis cracks intersect the same bedding surfaces as some of the studied specimens. Also common in this deposit are '*Spirophyton*', 'bromalites', and 'diminutive plug-shaped burrows'. *Undichna* co-occurs on the same bedding surface with the last of these.

Material– **WF**: Type 1: AM 7618 (two specimens), uncollected (one specimen); Type 2: uncollected slab (one specimen described based on field photographs); Type 3: AM 7618 (one specimen), AM 7619 (one specimen)

Description– Shallow, bedding-parallel epichnial grooves that are 2–6 mm in diameter and less than 1 mm deep. The grooves form an open curve or are, rarely, straight. They form sinusoidal waves only in the more complete specimens (AM 7618, AM 7619). Where measurable, they are 300–380 mm in wavelength with an amplitude of 95–240 mm. Parallel curvature between some of the grooves is suggestive of pairing, with a spacing of 14–140 mm. Where the grooves cross each other, previously formed ones are deformed in the direction travelled by those formed later. Several factors inhibit convincing ichnospecific assignment of this material, including discontinuity of the drag marks, overlays of drags by multiple tracemakers, but most importantly the incompleteness of the material due to the large size of the traces. They are nonetheless placed in three morphotypes based on the more complete specimens.

Type 1 includes specimens that show a pair of sub-parallel, curved grooves of similar wavelength that are in phase and do not cross, 45–115 mm apart. The maximum length in the

sample is 303 mm. None of the specimens exhibit a full sinusoidal wave, but this is likely due to the limitations of rock slab size. All occur on the same bedding surfaces as more complex *Undichna*.

Type 2 is based on one specimen that was photographed in 1999 by Dr Gess, but was not collected (Fig. 6.21A). The trace is a pair of sub-parallel grooves, 85–185 millimetres apart, central to which is a single sinusoidal groove. The paired grooves are approximately straight, both exhibiting a single, slightly out of phase curve in the same direction. The central sinusoidal groove has a wavelength of approximately 370 mm, its direction of travel is slightly oblique to that of the paired grooves.

Type 3 is based on two specimens (Figs 6.20B and 6.20C). Both contain at least two sets of bifurcating, out of phase, sinusoidal grooves of similar wavelength, which interweave in a complex pattern. On AM 7619, such drags produce a wavelength of approximately 300 mm with an amplitude of 150 mm. Direction of travel is indicated by the deformation of the some of the marks by those that were made subsequently. The sinusoidal wave is slightly skewed in the direction of travel in both specimens of this type (AM 7618 and AM 7619). The specimen on AM 7618 has a wavelength of 380 mm and an amplitude of about 240 mm.

Remarks – Undichna ichnospecies were reviewed by Minter and Braddy (2006) and Trewin (2000). The ichnogenus records drag marks produced by the fins of swimming fish on the sediment surface. The present specimens are retained in open taxonomy because interpretation of the material requires a specialist. They likely represent at least two or three ichnospecies and a variety of tracemakers. Klug *et al.* (2021) described a large *Undichna britannica* from the uppermost Devonian (Hangenberg Sandstone equivalent) of Morocco. They reported the only Devonian *Undichna* from Gondwana, and the oldest vertebrate trace fossil from Africa. Based on a revised stratigraphy for the Waterloo Farm locality, these traces are therefore the oldest known *Undichna* from Gondwana. The Waterloo Farm

Undichna are unusually large for the ichnogenus, but are similarly sized to the Devonian specimen from Morocco, which were interpreted to have been produced by an acanthodian about 0.5 m long.

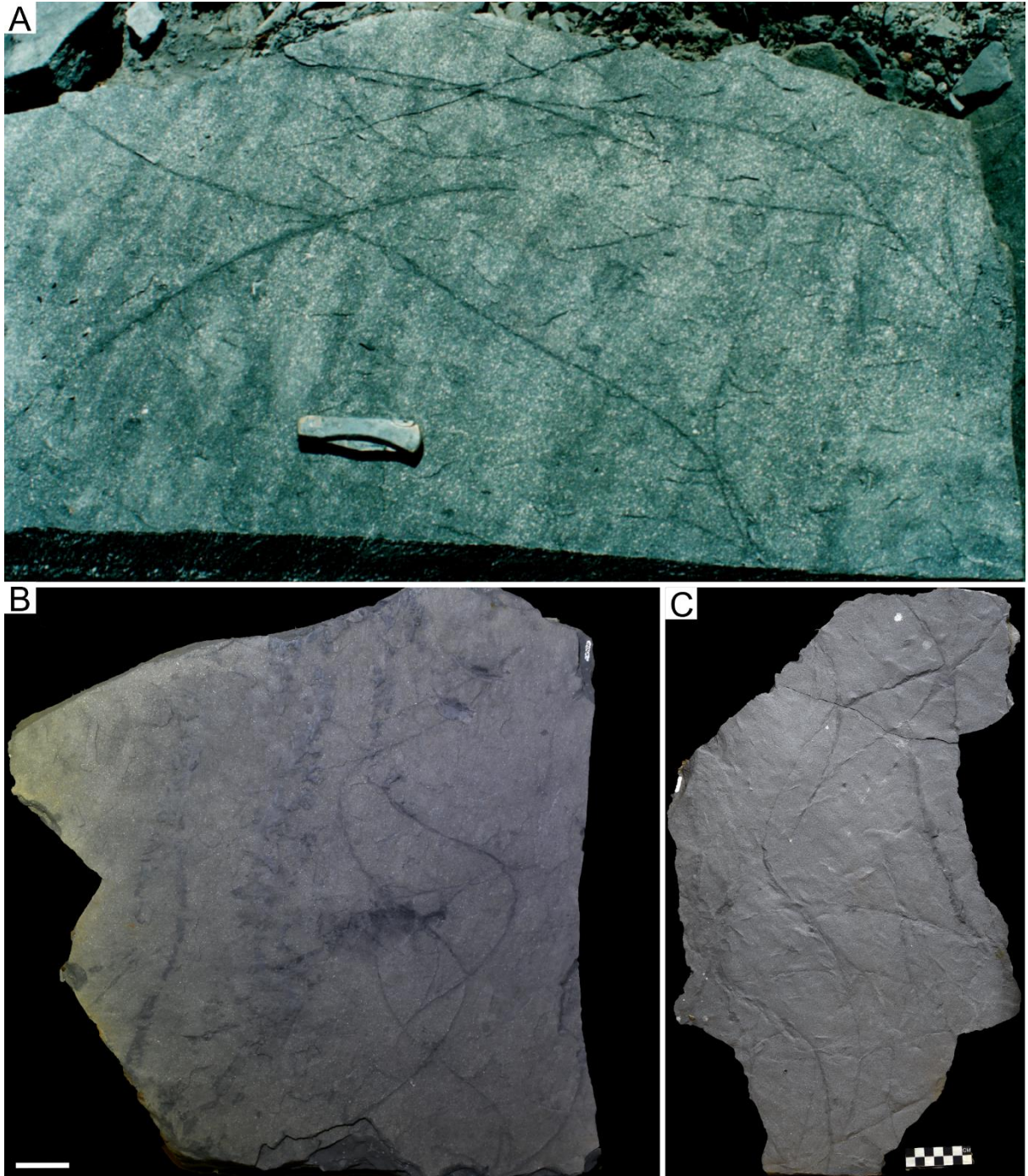


Figure 6.21: *Undichna* ispp. from the Witpoort Formation. A) Uncollected slab photographed at outcrop by R.W. Gess in 1999 (pocket knife is 7.2 cm long). B) (AM 7619, scale bar = 5 cm). C) (AM 7618, scale bar = 7 cm).

Ichnogenus *Zoophycos* Massolongo 1855*Zoophycos* isp. cf. *brianteus* Massolongo 1855

Fig. 6.22 A–F

Discussion – *Zoophycos* is widespread trace fossil, to which a wide variety of material has been assigned, and is sometimes regarded as a ‘*Zoophycos* Group’ rather than a well-defined ichnogenus. The material from Massolongo’s 1855 collection was studied recently (Olivero 2007). The *Zoophycos* type species had long since been synonymised with an extant alga (*Postelsiopsis*), and from amongst Massolongo’s remaining *Zoophycos* collection housed at the Natural History Museum of Verona, Italy, a lectotype ichnospecies was selected (Olivero 2007). *Zoophycos brianteus* (the lectotype ichnospecies) is a helical spreite structure, about 10–13 cm in diameter with a slightly lobate margin, which lacks a marginal/peripheral tunnel (Olivero 2007). The ethology of *Zoophycos* is debated; many researchers consider the trace to represent *in situ* deposit-feeding (e.g. Wetzel and Werner 1980; Olivero and Gaillard 1996, 2007; Seilacher 2007), but alternative models include subsurface disposal of faecal material (Kotake 1991), storing of nutrients derived from surface detritus in the subsurface (Löwemark 2015; Uchman and Wetzel 2024), and cultivation of microbes (Bromley 1991). The latter three models are, to a great extent, based on the downwards transport of sediment from overlying layers demonstrated in material from Pliocene deep-marine deposits (Kotake 1991). It is contended that no single model can account for all *Zoophycos*, particularly with the variety of material that has been assigned to the ichnogenus (Olivero and Gaillard 2007). *Zoophycos* occurs in marine strata from the Cambrian to Recent. In modern settings, and ranging back to the Mesozoic, *Zoophycos* is a typically deep-marine trace, however, in Palaeozoic strata the trace occurs in intertidal, shallow and deep-marine settings (Bottjer *et al.* 1988). The trace is typical of poorly oxygenated marine environments (MacEachern *et al.* 2010). Terminology for their description is according to that of Olivero and Gaillard (2007).

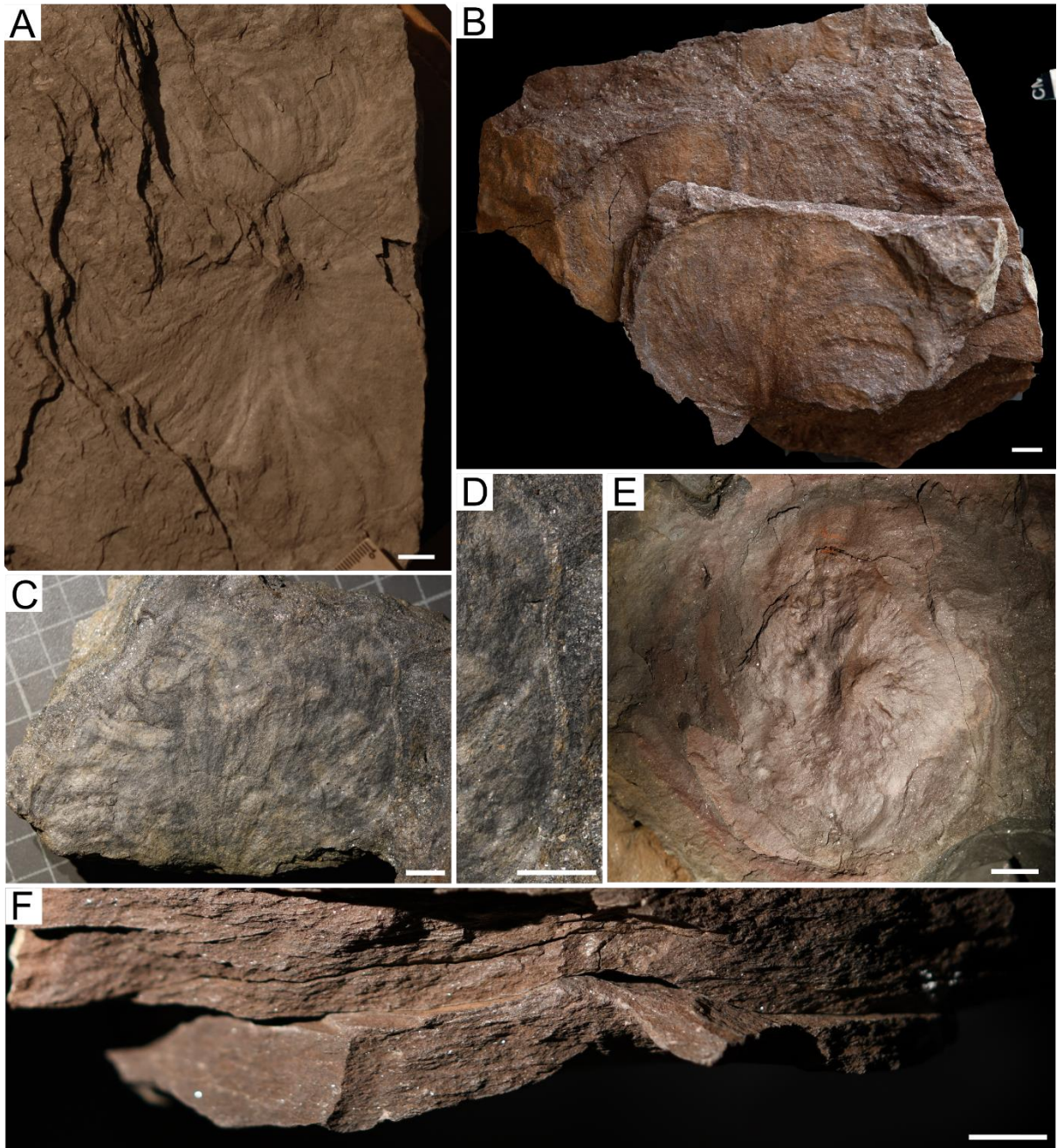


Figure 6.22: *Zoophycos* isp. from lower Witpoort Formation localities (all scale bars = 1 cm): A) Rabbit Ridge. B, E, F) Dam near Firglen turnoff (F is a profile view of the specimen in B). C, D) Howison's Poort. Specimens shown in B, D and E are counterparts. D represents an enlargement from C, showing a distinct clear peripheral tunnel that follows the slightly lobate margin of the trace. Photograph B shows a specimen that increases in diameter upwards. In photograph E, the spreite appears to have a pelletoidal texture superimposed.

Material – **HP**; AM 7643, AM 7644 a, b, AM 7645 a, b, AM 7646; **FG**; AM 7604 a, b, AM 7605, AM 7592 a–c, AM 7582; **RR**; AM 7783 a, b, AM 7784 a, b,

Occurrence – *Zoophycos* is only recorded in the lower Witpoort Formation in the study area (not including stratigraphically lower and higher occurrences of ‘*Spirophyton*’, which are described below). The trace fossil is superabundant and widespread at a black siltstone horizon about 12 m above the base of the Witpoort Formation at Howison’s Poort and at a similar level about 3 km to the east. A dam cutting nearby, ‘near Firglen turnoff’, preserves numerous *Zoophycos* in siltstone and overlying micaceous, medium-grained sandstone, from what appears to be a similar stratigraphic horizon. The trace is also recorded within a carbonaceous mudstone at Rabbit Ridge, about 90 m above the base of the formation, in low numbers.

Description – Helical structure comprising sub-horizontal spreite (lamellae) constructed radially in vertically-stacked whorls (laminae) about a central vertical or sub-vertical axis. The lamellae are arcuate or S-shaped, 1–4 mm wide, extending from a slightly raised central axis outwards to the margin. In some specimens the lamellae have a faint transverse ornamentation. The central axis forms a peak about which the trace is built. In no specimen was this axial peak found to be connected to an overlying horizon via a tunnel. The lamina is a roughly circular structure (69.5–>300 mm in diameter) in bedding plane view, with a lobate outline defined by a series of overlapping petallate lobes. It is usually distinct in colour from the matrix, and contains a higher proportion of mica. The margin of the lamina is usually bordered by a continuous or discontinuous peripheral tunnel of similar or slightly greater diameter than the lamellae. Laminae are stacked vertically to a maximum observed depth of 25 mm (though true depth may have been considerably higher). The sample set includes specimens that increase in diameter upwards (Fig. 6.22B), which is unusual for the ichnogenus.

Remarks– This material is consistent with a ‘strip mining’ construction model, whereby each lamina is worked once and is not revisited (Seilacher 2007). There is no evidence in the

studied sample for reworking of a trace in the given sample, nor do adjacent burrows intersect, and this is inconsistent with a farming trace (agrichnion) (Bromley 1991). Nor is there evidence of reworking in Recent or ancient *Zoophycos* (Löwemark 2015). Like the type species, *Zoophycos brianteus*, the described trace is a helical spreite structure enveloping a central axis, with a slightly lobate margin (Olivero 2007). It differs, however, in the presence of a peripheral tunnel, which is not preserved in the lectotype, but is a common and variably occurring feature of the ichnogenus (Miller 1991). Based on the axial peak rather than an axial tunnel, some of the Witpoort Formation material must have been constructed from the bottom upwards. At each successive turn of the helix the former connection of the underlying whorl to surface was truncated. Both upwards and downwards construction of *Zoophycos* traces has been suggested (Olivero and Gaillard 2007).

Zoophycos form ‘*Spirophyton*’ Hall 1863

Fig. 6.23 A–E; 6.24 A–F

Discussion – *Spirophyton* is a typically shallow-marine trace fossil in Palaeozoic strata (Hantzschel 1975; Miller and Johnson 1981; Kotacke 1991), however, its ichnogenic separation distinction from *Zoophycos* is controversial. Comparing *Spirophyton* in its type area, the upper Devonian of New York State, with *Zoophycos*, Miller and Johnson (1981, p. 1016) suggested that, “Although *Spirophyton* is smaller than *Zoophycos*, the broad morphologic similarity indicates that the two ichnogenera reflect generally similar behavior”. These authors nevertheless followed Simpson (1970) in retaining *Spirophyton* for helical *Zoophycos*-like structures that are circular in plan view and less than 8 cm in diameter (Miller and Johnson 1981; Miller 1991). The absence of a marginal/peripheral tunnel has been suggested as a further characteristic distinguishing *Spirophyton* from *Zoophycos* (Hantzschel 1975; Kotake 1991), but this feature is variably expressed in both ichnogenera (Miller 1991). Kotake (1991) suggested that upward recurvature of the spreite (J-shaped in profile) is

characteristic of *Spirophyton*, separating it from the more planar spreite of *Zoophycos*, but recorded both *Spirophyton* and *Zoophycos*-like morphologies in the same burrow systems. The ichnogenic status of *Spirophyton* is therefore dubious (Olivero 2007; Seilacher 2007), given that the main criterion for their distinction, size, is an unreliable ichnotaxobase (Bertling *et al.* 2006).

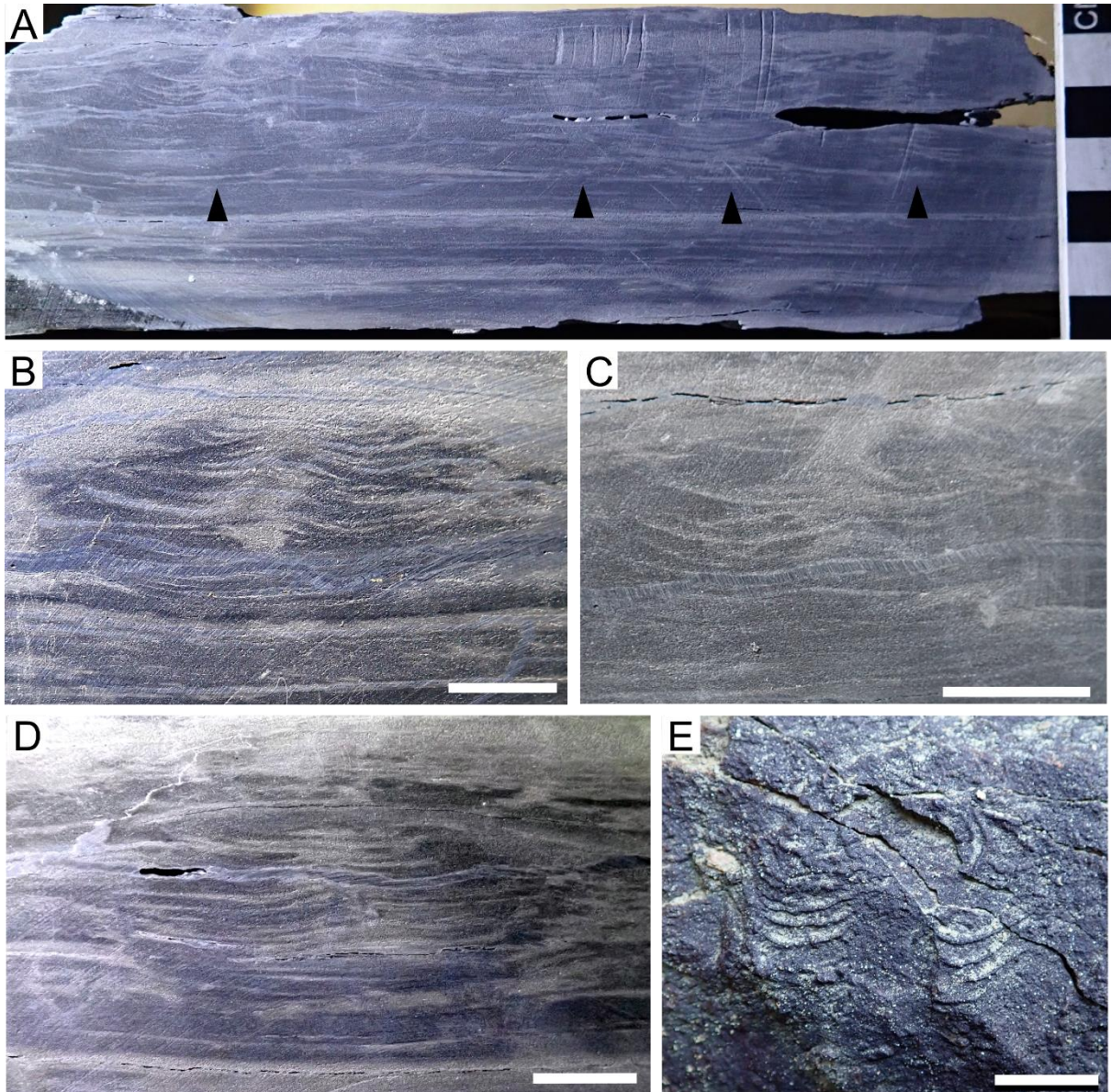


Figure 6.23: *Spirophyton* burrows in profile (scale bars for B–E represent 10 mm). A) Cut block of Waterloo Farm MFL black shale (AM 7612) exposing four burrows (arrows) which all occur at a similar level in the upper half of the exposed profile. B–D) Three examples from the block shown in A. E) An example from the upper Weltevrede Formation at Howison's Poort.

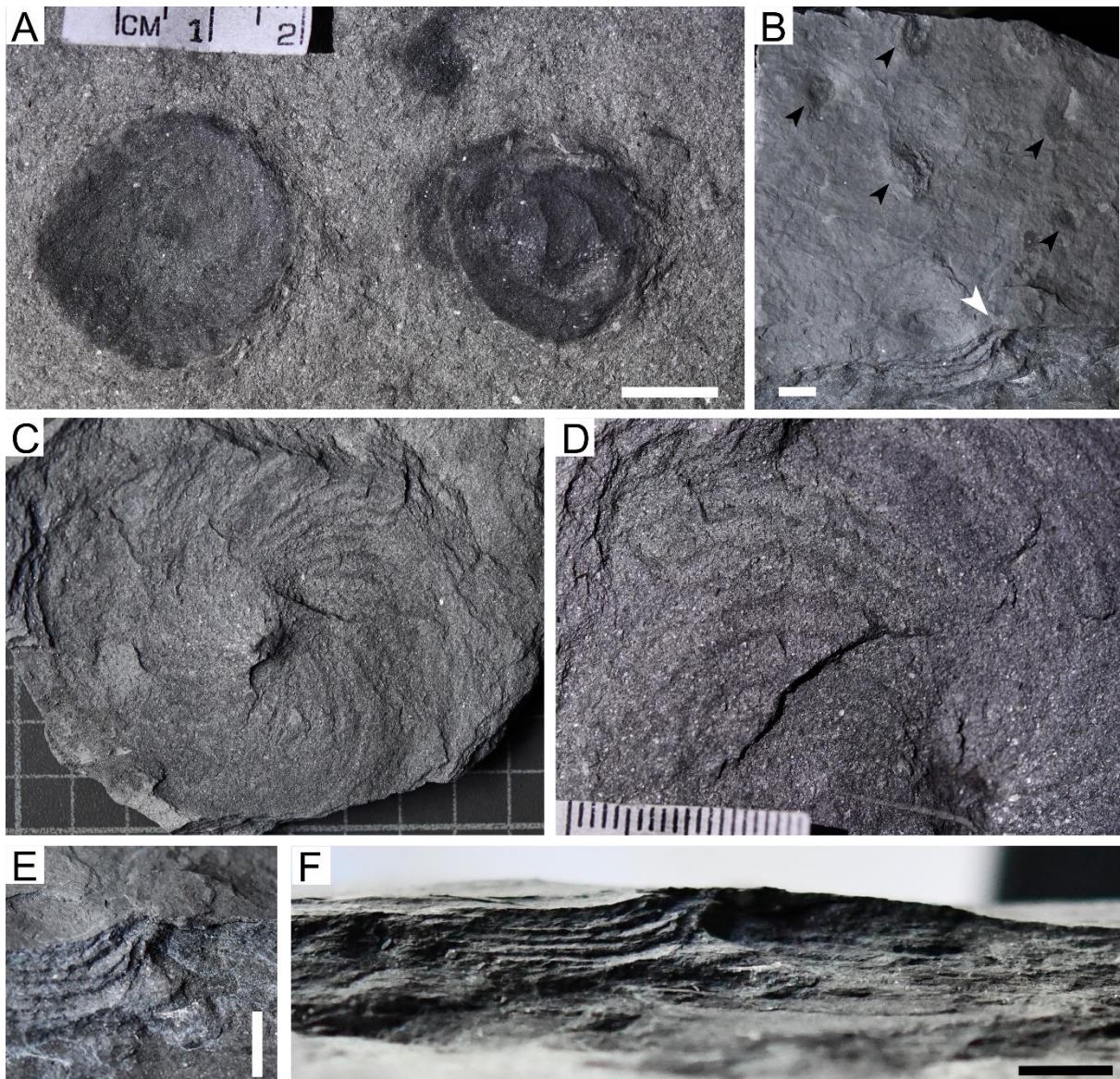


Figure 6.24: *Spirophyton* from the Waterloo Farm locality (all scale bars represent 10 mm unless otherwise stated). A) Two adjacent burrows (AM 7607), note the concentric appearance of the burrow on the right resulting from the truncation of several spreite. To the top of the photograph is an isolated axial tunnel. B) Several isolated axial tunnels are visible on a bedding surface on AM 7652a (black arrows). The burrow near the bottom of the photograph (white arrow) is exposed in profile and is enlarged in E and F. C) A well preserved spreiten showing primary lamellae (AM 7680a) preserved as darker and lighter bands (scale increments represent 1 cm) D) Counterpart to specimen shown in C, illustrating the primary lamellae, secondary lamellae are visible as subtle transverse ornamentation within the primary lamellae (scale in mm). E) Oblique view of burrow exposed in profile (AM 7652a) showing inverse cone-in-cone morphology of axial tunnel. F) Vertical profile of burrow shown in E.

Material– **WF**: AM 7653, AM 7652 a, b i–iii (three specimens), AM 7635 (three specimens), AM 7614, AM 7612 a, b (18 specimens), AM 7607 a, b, c (20 specimens), AM 7608 (six specimens), AM 7680, AM 7681, AM 5267, AM 5292, AM 5404 (two specimens), AM 5266

Occurrence– Superabundant *Spirophyton* occur towards the upper part of a ≤ 2 m thick black shale bed, the MFL at Waterloo Farm (Gess and Hiller 1995a). Although burrows may occur in high density, adjacent burrows do not overlap. The form is also recorded in the upper Weltevrede Formation at Howison’s Poort but is not abundant there.

Description– Helical structure comprising sub-horizontal spreiten constructed radially about a central vertical or sub-vertical axis. Up to 20 concave-upwards laminae are stacked vertically at 1–2 mm intervals, each about 0.5 mm thick (Fig. 23A–E). The diameter decreases upwards slightly. In the bedding plane expression, a spreite has a continuous sub-circular margin, with arcuate lamellae (2–3 mm thick) extending from central axis to the margin, turning upwards slightly towards the margins and axial peak. The lamellae curve in a sinistral or dextral direction with respect to the axis. Rarely the lamellae contain a faint transverse texture, particularly in specimens preserved in silt-rich black shale matrix. No peripheral tunnel is observed bordering the lamina. Diameter is 14–85 mm ($n = 58$, $\bar{x} = 31.5$ mm). Depth of penetration in full profile exposures is 20–35 mm ($n = 7$). The spreite do not differ in grain size from the dark grey shale matrix, but may be slightly darker or lighter in colour. A concentric pattern seen on some of the specimens is due to the truncation of successive bedding oblique laminae by splitting of the rock along shaly partings (Fig. 6.24A). Similar truncation of the axial peak likewise produces a circular structure, sometimes with slight grain size differences, representing underlying material that has been deflected upwards (Fig. 6.24A, C and E). Isolated axial tunnels overlying *Spirophyton* traces (AM 7652a) intersect a bedding plane probably representing the burrow openings to the sediment surface

(Fig. 6.24B). They are oval shaped and are up to 12 mm in long diameter, significantly wider than the axial peaks central to the spreite.

Remarks– The lack of a peripheral tunnel, approximately circular margin and small size of the traces is consistent with the notion of *Spirophyton* as a distinct form from *Zoophycos* cf. *brianteus* in the study area. These forms have only a slight overlap in diameter (14–85 mm vs. 69.5–>300 mm). Aside from these morphological and size differences, the activities represented by the two forms is very similar. Both are interpreted as systematic deposit-feeding structures constructed from the base upwards. According to Pemberton *et al.* 1988 (p. 866) “morphological features resulting from behaviour at a high level of significance are used for classification at the ichnogenic level”. Based on the Witpoort Formation material it would be difficult to argue that the differences between *Spirophyton* and *Zoophycos* are significant, in agreement with their suggested synonymy (Olivero 2007; Seilacher 2007).

Form ‘*Spirophyton*’ does not appear to co-occur with *Zoophycos* in the study area. The morphological differences and paucity of overlap in the same deposits has been explained by competition and exclusion between closely related species (Miller and Johnson 1981). An alternative prospect is that environmental constraints have led to the observed morphological differences. Diminution and behavioural modification of the *Zoophycos* animal in response to environmental stresses is a plausible alternative explanation for the apparent non-cohabitation.

Incertae sedis

‘Plug-shaped burrows’

Fig. 6.25 A–E

Material – **CH**; AM 7740 (two specimens); AM 7741 (two specimens); AM 7749 ai (one specimen); AM 7749 aii (two specimens); AM 7749d (one specimen); **WF**; AM 7611 (one

specimen); AM 7615 (one specimen); WF section 1a (uncollected, seven specimens); WF section 2c, m 4.5 (uncollected, three specimens)

Occurrence— Occurring in low to moderate abundance with low diversity to monospecific horizons in heterolithic facies. Associated with *Ptychoplasma*, *Scalichnus* and *Archaeonassa* in Facies Sc underlying Inner Outcrop mudstone horizon at Coombs Hill (Fig. 6.25B) and occurs in sandstone underlying MFL at Waterloo Farm (Fig. 6.25A), as well as elsewhere at the locality (monospecific horizons from WF 1 and 2; Fig. 6.25C–E). Also recorded in the upper Weltevrede Formation at Howison’s Poort.

Description— Vertically-orientated, plug-shaped burrows with a slightly asymmetrical, hemispherical, bowl-shaped profile, slightly flattened at the base. The walls are smooth, tapering downwards, sometimes with a thin lining. Burrow fill is structureless or contains convex-upwards laminae and is either similar in texture or coarser grained compared to the host rock. In bedding-parallel plan view the trace is sub-circular to sub-oval in shape, commonly with an irregular concentric or spiral fill. Diameter is usually between 15–25 mm. Width is greater than the depth of the trace, reaching up to 2.5 times the depth, though the apparent depth may be increased by retrusive cone-in-cone stacking. Retrusive structures are common, with a maximum recorded depth of 38.5 mm.

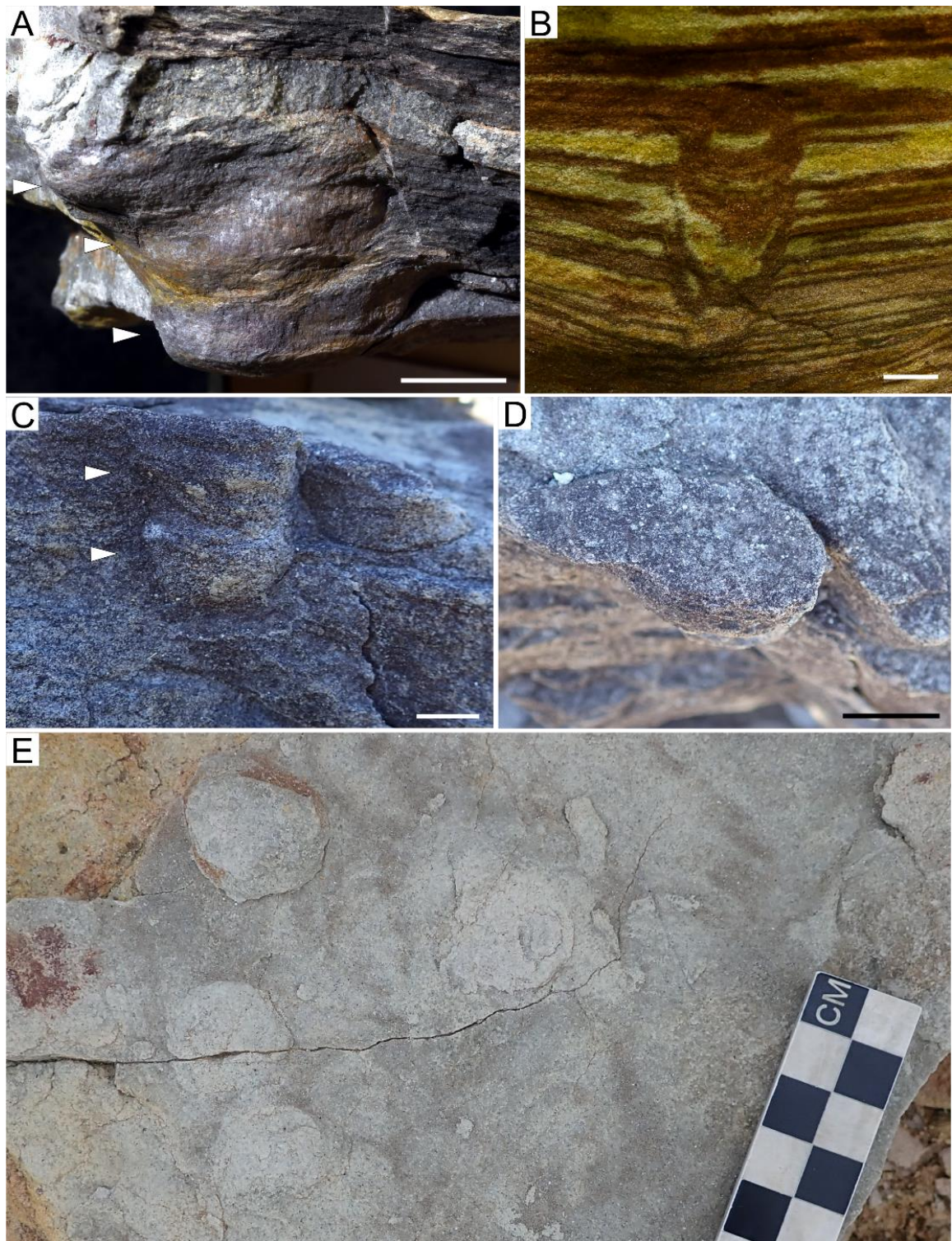


Figure 6.25: ‘Plug-shaped burrows’ from the Witpoort Formation (all scale bars = 10 mm). A) Retrusive structure with three vertically-stacked burrows (arrowed) from Waterloo Farm (AM 7611). B) Retrusive burrow from Coombs Hill (AM 7740). C) Retrusive cone-in-cone structure from the upper Witpoort Formation near Waterloo Farm with two burrows vertically stacked (arrows). D) Bedding plane expression of the burrow shown in C. E) Several hypichnial burrows from a channel fill succession underlying the Waterloo Farm fossil locality.

Remarks – Isolated and retrusively stacked structures that share a similar, bowl-shaped morphology can be considered congeneric because upwards migration of the tracemaker evidenced by the retrusive structures is a response to rate of sedimentation, and this does not provide a reliable ichnotaxobase (Bromley 1990). Absence or presence of meniscae within the plugs is indicative of rate of upward migration. Morphologically the structures are comparable with single opening, plug-shaped burrows, which comprised eight ichnogenera, usually ascribed to sea anemones (Pemberton *et al.* 1988). Though sharing a degree of morphological similarity with members of the group (*Conichnus*, *Conostichus* and *Bergaueria*), a slight asymmetry in the Witpoort Formation traces distinguishes them from the group, which have radial symmetry, as characterised by Pemberton *et al.* (1988). The material, however, presents a strong similarity to vertical equilibrium structures from fluvial deposits in the Frasnian (Late Devonian) of New York State, which are closely associated with non-marine unionid bivalves, their inferred tracemaker (Bridge *et al.* 1986). They share the retrusively-stacked, slightly asymmetrical plugs and the irregular oval-shaped cross section. Indeed, bivalve-generated equilibrium structures present a similar set of morphologies to the actinian-derived group of structures and may not always be clearly distinguishable (Zonneveld and Gingras 2012). Some of these traces may represent partial *Scalichnus* (with emergent vertical siphon traces not visible or truncated).

‘Diminutive plug-shaped burrows’

Fig. 6.26 A–C

Material– **WF**; AM 7636 (tens of specimens); AM 18039 (10’s of specimens); AM 18040 (six specimens); AM 18041 (10’s of specimens); AM 18042 (10’s of specimens)

Occurrence– Abundant trace in the MFL black shale. It usually occurs in dense accumulations. Two slabs (AM 18039, AM 18042) contain *Undichna* trails on the same

horizons. In one instance the fish trail cross-cuts and deforms a burrow, indicating that the burrow was formed previously and was probably open to the sediment surface.

Description– Diminutive, vertically-orientated, short, plug-shaped burrows with sub-vertical, downwards-tapering walls and a flat base (Fig. 6.26B). Measuring 1.5–9 mm in diameter, up to 3 mm deep, wider than deep, comprising a core of darker or lighter coloured sediment around half of burrow diameter, surrounded by an outer ring with radial striations (Fig. 6.26C).

‘Micaceous backfill structures’

Fig. 6.26 D

Material– **WF**: AM 18043 (three specimens); AM 18044 (two specimens); AM 18045 (one specimen)

Occurrence– The trace is recorded in laminated black shale from the Waterloo Farm MFL deposit. No other traces are recorded on the slabs containing this trace.

Description– Unlined horizontal burrows with indistinct margins containing subtle transverse arcuate backfill laminae. The specimens have open curves and sharp (90°) turns. Burrow fill is enriched in mica compared with the host rock. Diameter is 2.5–9.5 mm, with individual specimens varying by about 30% in diameter. Length ranges from 11.5 to 60 mm in the studied sample. One specimen cross-cuts another (AM 18044).



Figure 6.26: Enigmatic traces from Waterloo Farm MFL (scale bars and increments = 1 cm). A–C) Diminutive plug-shaped burrows. D) Micaceous backfill structures.

‘Compound trace 1’

Fig. 6.27 A–F; Fig. 6.28 A–D

Material– Witpoort Formation: **RR**: AM 7792 (R1 030); AM 7787 a + b (R1 033); AM 7789 (R1 ss 026); AM 7790 (R1 026); AM 7785 (a + b), AM 7799 a + b; AM 7800; **WF**; AM 7626; AM 7606; AM 7615; AM 7625; AM 7626; **SF**: uncollected material from various horizons

Weltevrede Formation: **HP 1**: (AM 7670, 7673-5)

Occurrence– Occurs in parallel-laminated, thin-bedded fine-grained sandstone facies. At Rabbit Ridge (Upper Section) in low density, commonly observed on bedding surfaces as diminutive vertical and horizontal tubules occurring in clusters. Strongly associated with *Nereites*. At Waterloo Farm the trace is pervasive as hypichnia on certain sandstone bed soles underlying MFL, and is associated with *Cruziana/Rusophycus*.

Description– A compound trace fossil comprising diminutive, irregularly-radiating, cylindrical tubes (1.5–3.5 mm in diameter) that branch horizontally and vertically, forming spreite. The tubes are endichnial structures expressed as irregular radiating networks on bedding surfaces, and in some hypichnial examples, are observed in connection to short, sub-horizontal burrows (Fig. 6.27 A–F).

Spreiten comprise networks of small cylinders expressed on bedding surfaces in roughly circular, vertically-orientated clusters (up to 53 mm in diameter) or as horizontal feathery spreite structures that fan out discontinuously along the bedding planes. Both vertical and horizontal spreite co-occur on bedding surfaces (Fig 6.28A). Cylinder fill is apparently structureless, usually similar in texture to the matrix. Occasionally the tubes are slightly darker in colour than the host rock.

Where bedding-parallel the cylinders are straight to gently curved, up to 18 mm long, sometimes occurring as several adjacent, parallel spreite-like structures. The tubes often have

a segmented appearance, sometimes suggesting an apparent helical disposition. This may result from repeated upward turns of the tubes. In hyporelief the tubes generally weld into the overlying bed surface, apparently turning upwards at their terminations (Fig. 6.27C, D). In negative epireliefs on draped sandstone bed tops the tubes are bordered by shallow pushup ridges (Fig. 6.27E). In this expression the tubes appear to turn either upwards or downwards where diverging from the bedding surface of exposure.

Three-dimensional ramification of a vertically-disposed cluster of tubes was assessed by parting a specimen at three equally spaced levels in an 8 mm thick profile. The tubes originate at the lowest level with a tight, 6 mm wide cluster of six closely spaced bedding-oblique tubes (Fig. 6.28D). The middle horizon (Fig. 6.28C) shows that the two tight clusters from the lower level have expanded upwards into broader, roughly circular clusters (about 1–2 cm in diameter), comprising eight and five tubes respectively. At the highest level the tubes turn bedding parallel in a dense overlapping mass about 35 mm in diameter (Fig. 6.28B). An upward-branching, cone-shaped composite structure is therefore indicated.

Examples from Rabbit Ridge, Waterloo Farm and Seven Fountains show the spreite emerging laterally from short, amorphous or segmented horizontal burrows (Fig. 6.27A–F). The originating horizontal structure is a smooth hypichnial ridge from which the spreite extend laterally on both sides for 15–30 mm along the bedding surface, turning upwards distally. The spreite extend asymmetrically in an oblique orientation to the central structure. Two burrows from Waterloo Farm (AM 7626) show a central segmented structure (Fig. 6.27C and D). The segmented structure in one contains transverse spreite resembling the tubes in shape and size, the other has larger V-shaped segments. Both amorphous and segmented central burrows are described based on very small sample sets, yet they nonetheless provide clear evidence of interconnection between the central locomotory structure and emergent clusters of spreite.

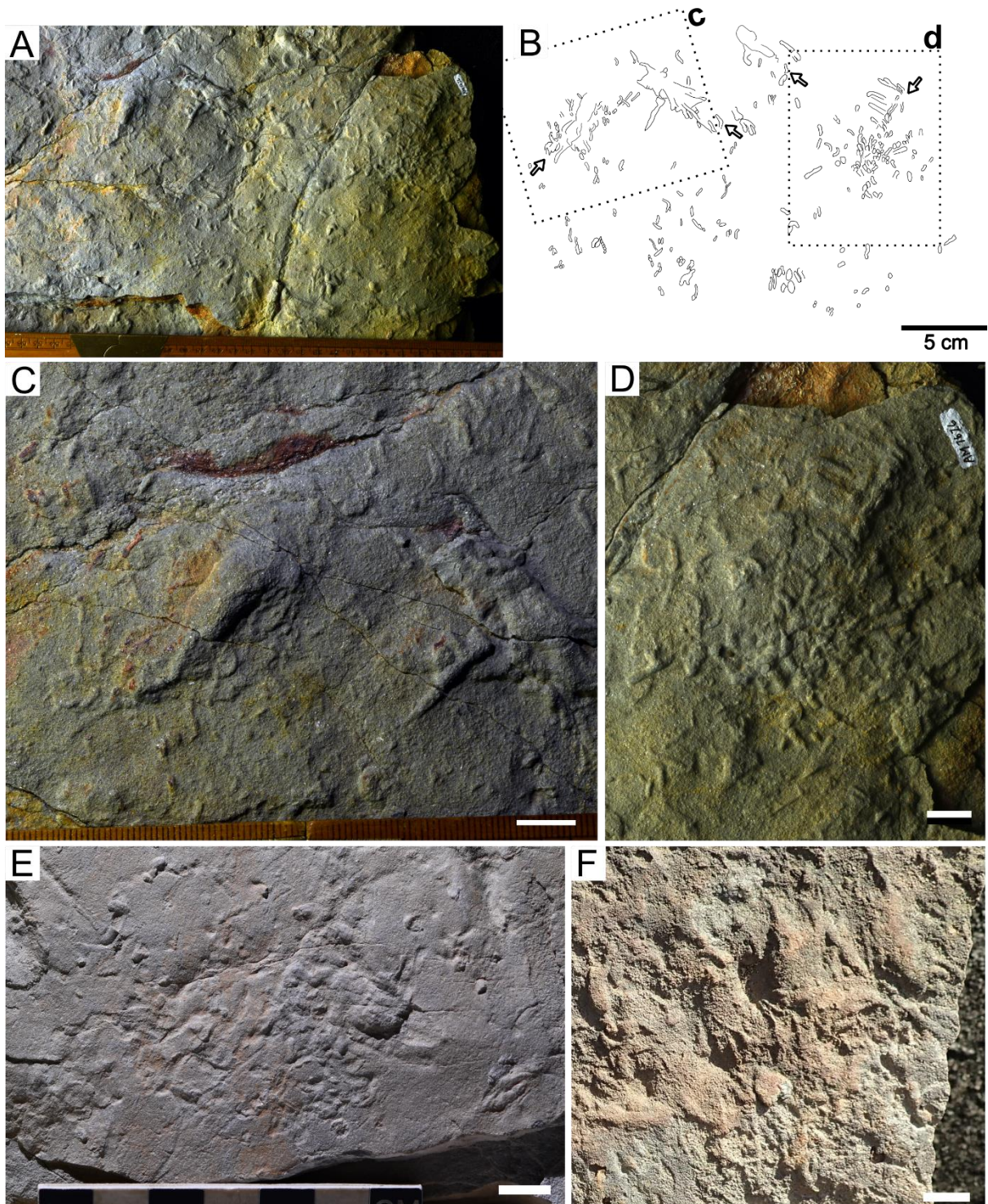


Figure 6.27: Compound trace 1. A) Bottom of slab from Waterloo Farm (AM 7626) and corresponding line drawing B. The areas enlarged in C and D are outlined on the drawing. Arrows indicate locomotory structures attached to radiating tubes. E) Top of bed from Rabbit Ridge with tubes radiating from what appears to be an underlying locomotory structure (AM 7790). F) Bottom of slab photographed at outcrop at Seven Fountains. Scale bars for C–F = 1 cm.

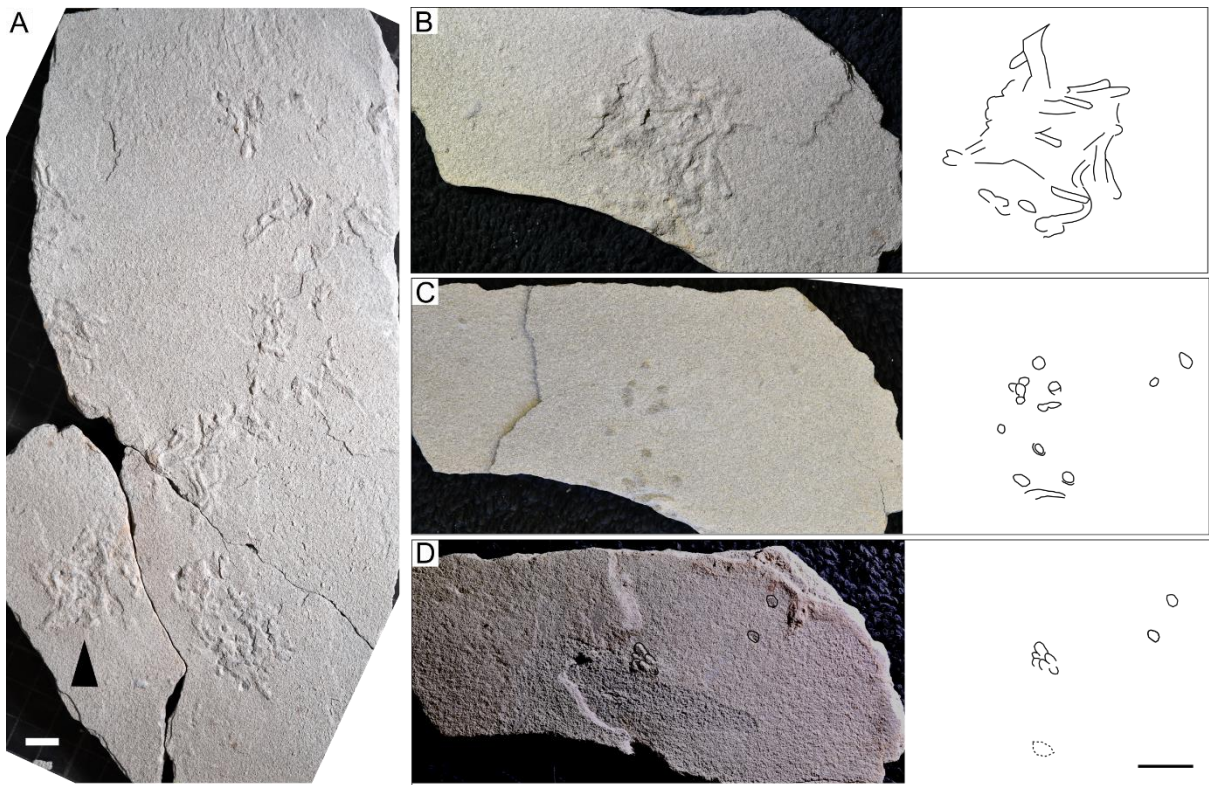


Figure 6.28: Compound trace 1 from Rabbit Ridge (AM 7785) (all scale bars = 1 cm, B–D to same scale). A) Tubes intersect the bedding surface in vertical clusters and horizontal feathery spreiten. Arrow indicates the trace enlarged in B. B–D) The trace is exposed at three levels at approx. 4 mm intervals from the lowest level D to the highest level B.

Remarks– This compound trace fossil appears to have elements of semi-systematic, selective working of the sediment combined with locomotion. This combination of structures is consistent with a motile deposit-feeding trace. Tubes that form a three-dimensional, upward-branching, cone-shaped cluster, as demonstrated in Figure 6.28, show a resemblance to *Parahaentzschelinia*, which has been attributed to deposit-feeding tellenid bivalves (Wikander 1980; Knaust 2015) or polychaetes (Reynolds and McIlroy 2017). Origination and radiation of the spreite from a central, sub-horizontal locomotory structure is more consistent with those expected for bivalvian siphonal-feeding traces than the spreite of vagile worms (Bromley *et al.* 2003).

Vagile, deposit feeding bivalves are responsible for a variety of compound traces in the rock record. ‘Compound trace 1’ differs from siphonal feeding traces such as *Arenituba verso*

or *Saronichnus abeli*, which are inferred as belonging to sessile infaunal bivalves (Pervesler and Zuschin 2004; Knaust 2015). *Bromleyia magnifica* is a compound ichnofossil with a chevronate locomotion trace (*Protovirgularia*) leading into intermittent resting traces (*Lockeia*) surrounded by radiating, bedding-parallel deposit-feeding spreite (*Lophoctenium*-like) emplaced symmetrically on either side of the resting trace, which is attributed to the activities of an infaunal cleft-foot bivalve (Ekdale and Bromley 2001; López Cabrera *et al.* 2023). Similar traces, illustrated by Poschmann and Braddy (2010, their Fig. 7), from the Early Devonian of the Rhenish Mountains, Germany, were interpreted as bivalve deposit-feeding traces. The Witpoort Formation material differs in the three-dimensionality of its spreite and in not being associated with either *Lockeia* or *Protovirgularia*.

Hillichnus lobosensis is a very complex, compound ichnotaxon from deep-marine deposits in the Palaeocene of California, interpreted as a deep-tier, deposit-feeding trace of a tellinacean bivalve (Bromley *et al.* 2003). The trace had long remained enigmatic in museum collections, but its full three-dimensional morphology has been thoroughly investigated (Bromley *et al.* 2003). It comprises a locomotory structure that is obliterated by the deposit-feeding activities, which form feather-like spreite extending horizontally on either side. Sub-vertical, curved tubes extend upwards at increments from the locomotion structure which connected to the former sediment surface via a vertical siphonate tube. *Hillichnus* provides a model that explains the overall morphology of ‘compound trace 1’, which differs in having a far more variable expression, particularly in the less systematic, three-dimensional ramifications of the spreite.

The variable activities recorded by this trace fossil could be consistent with those of a deposit-feeding tellenid bivalve, which exhibit remarkable variation in behaviour in modern intertidal and deep-sea environments (Bradfield and Newell 1961; Hughes 1969; Wikander 1980; Bender and Davis 1984). However, deposit-feeding tellenids are not considered to

predate the Cretaceous Period (Buatois *et al.* 2017). *Jamesonichnites*, a Jurassic-aged trace, indicates deposit-feeding tellenid behaviours in pre-Cretaceous strata (Bromley *et al.* 2003). It would be controversial, based on present evidence, to infer deposit-feeding tellenids in the Devonian Period.

‘Compound trace 2’

Fig. 6.29 A–C

Occurrence– ‘Compound trace 2’ is recognised at various horizons at Waaipoort including in the Blinkberg Formation (middle Weltevrede Subgroup), the uppermost Weltevrede Formation, the lowermost bed of the Witpoort Formation, and in the Skitterykloof Member. In all these examples it occurs in apparently monospecific assemblages with variable bioturbation intensities (BI 1–4), developed in heterolithic sandstone. It is also found in the Upper Section at Rabbit Ridge (lower Witpoort Formation) in similar facies immediately underlying a thick shale horizon. It occurs in one bed in low abundance (BI 0–1) overlying *Nereites* ispp. and is expressed on the same bedding horizon as ‘compound trace 1’.

Material– **RR**: AM 7792 (two specimens); **WP**: uncollected material from various horizons specified above.

Description– In vertical profile the trace comprises a central furrow from which branching tubes emerge symmetrically or slightly asymmetrically from the lateral margins. The tubes are 0.5–2 mm in diameter, extending horizontally from the central axis and may recurve upwards or downwards. Successive traces of this nature may be stacked vertically in some examples. The tubes ramify in variable ways and each occurrence shows slight differences that are described separately below. The central furrow is 1/6 to 1/10 of the entire trace diameter, which ranges from 25 to 135 mm in the measured sample.

In the Rabbit Ridge material a single bifurcation is visible at similar distances on either side of the central furrow. The two examples from this locality are relatively small (25 mm to 30 mm in diameter) with lateral tubes extending horizontally from a central furrow. One of them has lateral tubes that appear to bifurcate at about halfway along their length. The other specimen shows three successive lateral tubes stacked vertically. One of these traces is connected with the only bedding plane view of the trace observed in this study (Fig. 6.29C). It comprises an irregular central zone of reworked sediment with arcuate transverse ribs that are partially ferruginised, at intervals of 4–10 mm along the trace length. The trace continues along the bedding surface for about 6 cm and terminates in a circular structure resembling *Laevicyclus* with a diameter of 5 mm. The same bedding surface contains clusters of vertical to sub-vertical tubes with a similar diameter (compound trace 1) to which a connection is not established.

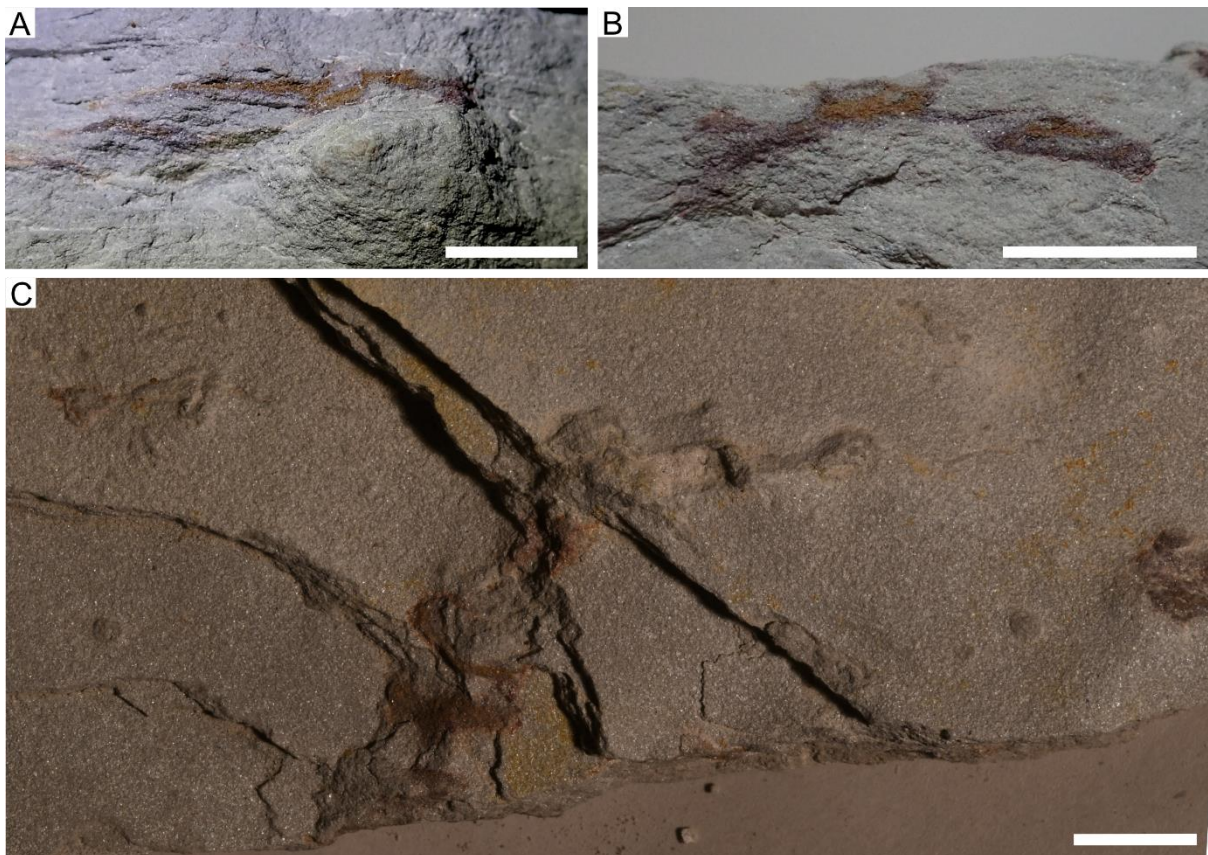


Figure 6.29: Compound trace 2 from Rabbit Ridge (AM 7792) (all scale bars = 1 cm). A–B) Two specimens observed in profile. C) The bedding parallel continuation of the trace shown in B.

Remarks– ‘Compound trace 2’ is not observed in connection to ‘compound trace 1’, but it represents a comparable set of behaviours, with a central locomotory structure that gives rise laterally to outwards and upward-branching tubes. It potentially represents a profile view of the trace. Understanding the full ramification and possible interconnection between these traces will require a concerted study. Both traces appear to be common in the Witpoort Formation. Larger and more elaborate examples of ‘compound trace 2’ than those illustrated were located at outcrop from multiple horizons at Waaipoort.

‘Bromalites’

Fig. 6.31 A–J

Discussion– Bromalites include fossilised digestive products (Hunt and Lucas 2012). They are described below following the terminology of McAllister (1996). Bromalites in the studied sample include material assigned to four morphotypes primarily based on shape, texture and presence or absence of inclusions. These include three coprolite forms and a putative regurgitalite form.

Occurrence– The majority (80%) of studied specimens are of the spiral coprolite form, which is abundant at Waterloo Farm (MFL, n=35) and two mudstone horizons at Coombs Hill (CH, n=26), and is less abundant in a prominent mudstone capping the Rabbit Ridge (RR, n=5) succession. The remaining two coprolite morphotypes (oval and dumbbell-shaped) and ‘regurgitalites’ are confined to small sample sets from MFL. The coprolites are sharp-walled structures that are less dense than the matrix. They are darker in colour than the shale matrix in fresh specimens, and appear to not have undergone significant compaction. The ‘regurgitalites’ are identified as thin films on bedding surfaces containing scattered inclusions, but with poorly defined boundaries. None of the studied bromalites are associated on the same bedding surfaces with other trace fossils.

Spiral coprolites

Material– (one specimen unless otherwise stated)– **MFL**: AM 5401 O; AM 5398; AM 5401 H; AM 5401 I; AM 5401 F; AM 5401E; AM 5401G; AM 5724; AM 5723; AM 18021; AM 18023; AM 18024; AM 18025; AM 18026; AM 18027; AM 18028; AM 18029; AM 18030; AM 18031 (three specimens); AM 18032; AM 18034; AM 18035; AM 18036; AM 18037 (two specimens); AM 7629; AM 7632; AM 7631; AM 7637; AM 7630; AM 7642; AM 7634; AM 7639; AM 7633; **CH**: AM 7729; AM 7737 (two specimens); AM 7738; AM 7730; AM 7722; AM 7721 (five specimens); AM 7720 (three specimens); AM 7723 (two specimens); AM 7724; AM 7725; AM 7731 (two specimens); AM 7733; AM 7750; AM 7735; AM 7734; AM 7732 (two specimens); AM 7727; AM 7728; **RR**: AM 7807, AM 7808; AM 7781 (two specimens). Variations to spiral form– **WF**: AM 18032 (one specimen), AM 18033 (one specimen); AM 18021; AM 5397 (one specimen); AM 5489 (one specimen)

Description– Organic-rich material forming a sharp-walled groundmass that is less dense than the rock matrix. Specimens attributed to this morphotype are spiral-shaped with the spiral axis orientated vertically. In plan view on bedding surfaces they are sub-circular or oval, with a slight lobe at the lateral margin, from there spiralling inwards towards the apex. The bottom surface is the flatter, whereas the coprolite tapers upwards to the apex. The apex (posterior end) is a central (or slightly off-centre) dome that is represented in counterparts by a bowl-shaped depression $\frac{1}{3}$ to $\frac{1}{2}$ of the trace diameter. In many specimens this apex has a granular texture. Though granules can occur sporadically in the remainder of the trace, they are concentrated, and are coarsest in this apical lobe. The granules are sub-spherical pellets of consolidated clay particles that are similar in composition to the remainder of the coprolitic mass. These ‘pellets’ are a small fraction of a millimetre to approximately 3 mm in diameter. The remainder of the groundmass is structureless. Specimens from MFL are generally darker in colour, slightly finer grained and depleted in mica with respect to the host rock. Weathered

specimens are similar in colour to the rock matrix, and this is the case for the CH and RR material, and part of the MFL material. In some of these samples a fine veneer of clay covers the coprolitic groundmass; an artefact only observed in weathered specimens. Some of the specimens from CH are stained with red, orange and yellow colours, and this extends into the matrix surrounding the coprolites.

Samples from the three localities show a significant overlap in diameter and similar mean diameters (Fig. 6.30). Size–frequency distributions from all three localities are bimodal. The depth of specimens is assessed only where profile views are attained or by removal of the coprolitic mass. One specimen from MFL is 5 mm deep and it is evident that all specimens have a lesser depth than those from CH, in which four specimens are 10.5–15 mm deep. The MFL sample set (n=35) has a diameter (measured on horizontal long axis) ranging between 5 and 35 mm (mean=15 mm). The Rabbit Ridge sample (n=5) is 8.5–22.5 mm in long axis diameter (mean=16.9 mm). The Coombs Hill sample set (n=28) ranges in diameter between 8 and 42.5 mm (mean=19.65 mm).

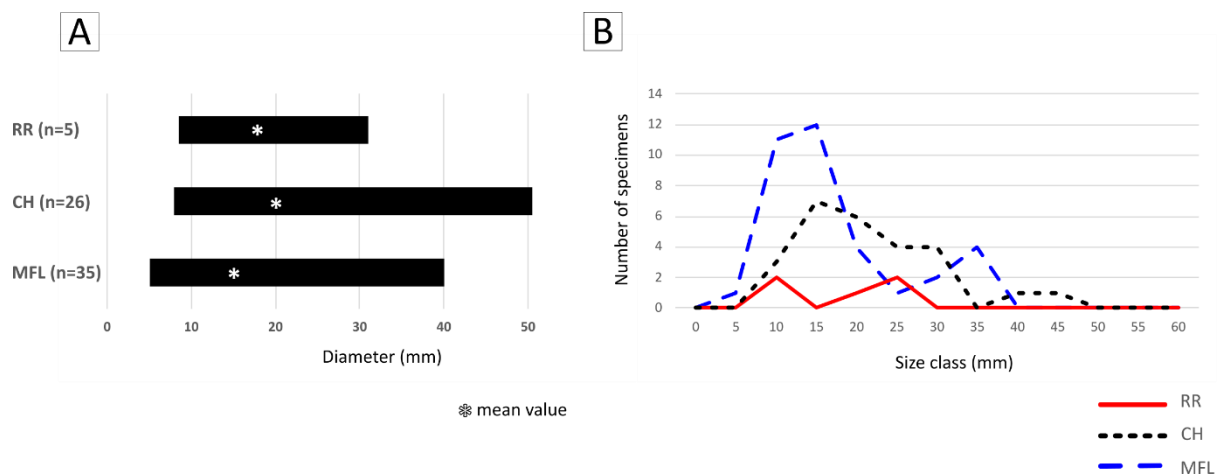


Figure 6.30: Size distribution of spiral coprolites from Rabbit Ridge (RR), Coombs Hill (CH) and Waterloo Farm (MFL). A) Size range plot. B) Size frequency plot.

Bivalve shells and putative shell fragments are the only recognisable inclusions in the spiral morphotype. From MFL, two specimens (AM 5401H and AM 5401E) have a bivalve shell partially underlying, extending onto the bedding surface adjacent to the specimen (Fig. 6.31B). In one of them, the length of the shell exceeds the diameter of the coprolite. Four specimens in the CH collection are underlain by intact bivalve shells (AM 7737; AM 7720; AM 7733; AM 7734) (Fig. 6.31C). The shells have their convex outer surface downwards in all cases. Several specimens contain angular fragments of a hard dark material that resembles the texture of bivalve shells from the site, which are the only fossils from CH that have this particular texture (Fig. 6.31E).

Variations to the spiral coprolite form exist in the MFL sample that show one or more spiral forms embedded in an amorphous groundmass, which in some examples resembles the oval coprolite morphotype. AM 5489 comprises a spiral form with two sub-circular, structureless forms adjacent. AM 5397 (Fig. 6.31F) is an 83 mm long structure containing two spiral forms embedded in an elongate structureless mantle. AM 18032 is a 37 mm long structure comprising a spiral with an elongate oval mantle. These latter two examples imply that both oval-shaped, structureless coprolites and spiral forms can emerge from the same tracemaker.

Remarks– Spiral coprolites are common at three localities in the study area. They have bimodal size-frequency distributions at all three sites. This may reflect two size classes of tracemaker. Spiral coprolites are formed by animals that have a spiral gut valve (Hunt and Lucas 2012). Based on the known fauna from Waterloo Farm, a variety of possible tracemakers can account for the spiral coprolites, including eurypterids (Hunt and Lucas 2012), placoderms, sharks and lungfish (Williams 1972), and coelacanths (Zaton *et al.* 2017). Coelacanths are unlikely trace producers because at this locality they are almost exclusively juveniles, and are too small to account for the coprolites. Phlyctaeniid arthrodire placoderms

(*Groenlandaspis* and *Africanaspis*) are the most abundant resident fish taxa from Waterloo Farm (Gess and Whitfield 2020). Their durophagous feeding habit would account for the bivalve shell inclusions that are common to the morphotype, and are considered likely producers, but eurypterid producers are not excluded.

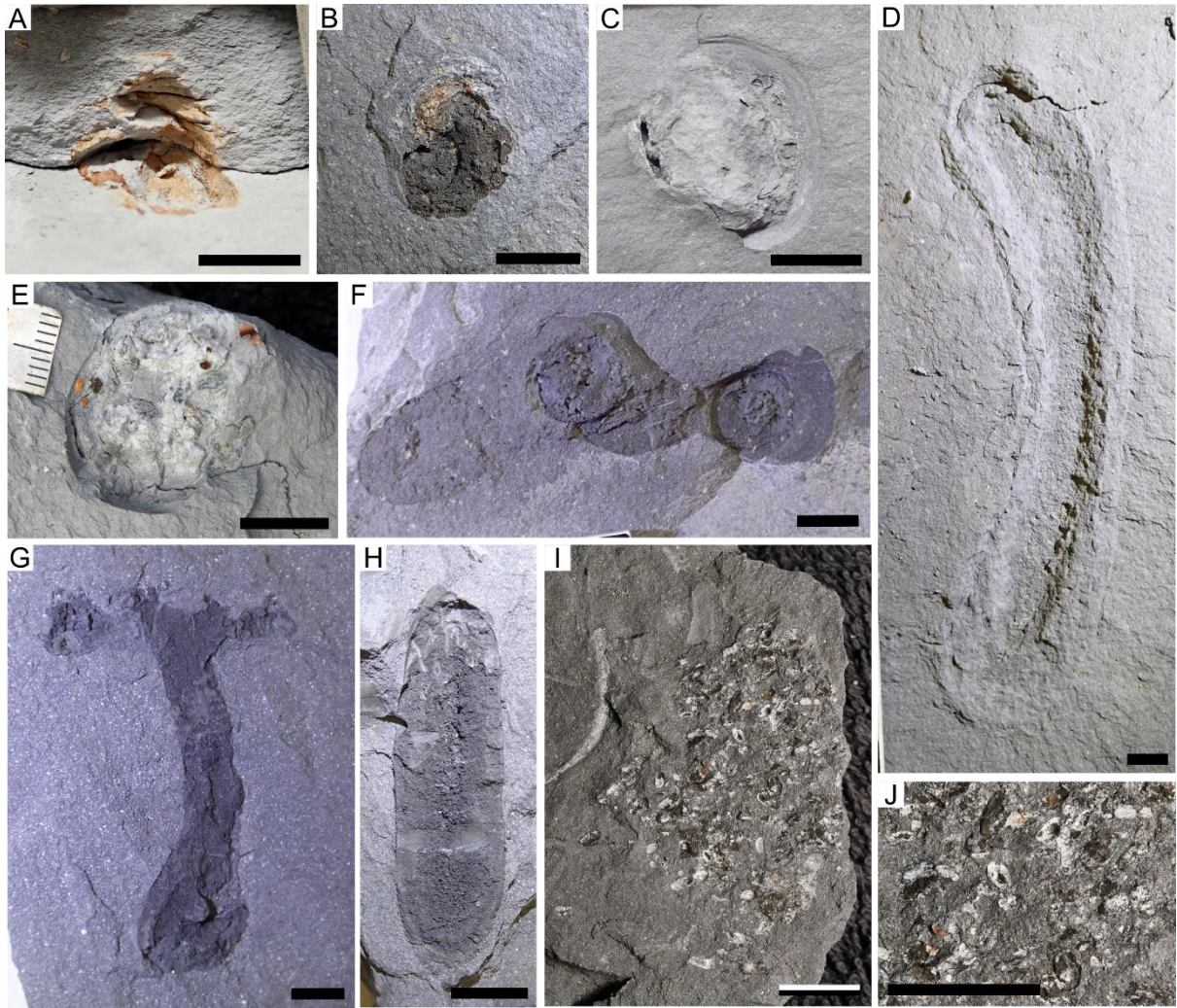


Figure 6.31: Bromalites from the Witpoort Formation (All scale bars = 1 cm). A) Spiral coprolite in profile from CH (AM 7734). B) Spiral coprolite (counterpart; AM 5401G) from MFL with bivalve shell superimposed. C) Spiral (?) coprolite from CH (AM 7733) underlain by bivalve shell. D) Elongate oval coprolite from MFL (AM 7624a). E) Spiral coprolite from CH (L3 066), with dark grey inclusions interpreted as bivalve shell fragments. F) A series of two spiral coprolites embedded in a groundmass from MFL (AM 5397b). G) Two superimposed, dumbbell shaped coprolites from MFL (AM 5725). H) Oval coprolite from MFL (AM 5735). I) Putative regurgitalite from MFL (AM 7622a). J) Enlargement from (I) showing scatter of miniscule, intact shells.

Oval coprolites

Material– WF: AM 18021 (two specimens); AM 5735; AM 7642

Description– Bedding-parallel oval structure containing structureless, dark grey, carbonaceous material. The outline is very even and unlobed. The length is 26–155 mm and varies from double to about five times the width. Thickness of the structure is difficult to measure, but apparently not more than a few millimetres thick. No fossil inclusions were identified. One of the specimens, the longest one, is slightly flared at both ends.

Two of the oval-shaped specimens (AM 5735, AM 7642) have an elongate ridge in the centre that is filled with coarser granules than the perimeter (Fig. 6.31D). In both cases this ridge is about half of the diameter and extends most of the length.

Dumbbell-shaped coprolites

Material– WF: AM 5725 (two specimens)

Description– Two associated, partly overlain specimens are 72 and 47 mm long, elongate structures, flared at both ends, with a narrow ‘neck’ in between that is about half as wide as the flared ends. The fill is a structureless, dark grey organic carbon rich matrix that is finer grained than the host rock. No inclusions are recognised in this morphotype.

Regurgitalites

Material– WF: AM 7621 a + b, AM 7622 a + b, AM 7623 a + b

Description– Regurgitalites in the studied sample are separated from coprolites in that they occur as an irregular scatter of material rather than a coherent mass, and contain various inclusions (Fig. 6.31I, J). They occur as thin films on bedding surfaces with indefinite boundaries, approximately 35–50 mm in diameter, but some of the inclusions are scattered beyond the perimeter of the main mass. They comprise dense accumulations of fossil material, particularly small, intact, disarticulated shells as well as other indeterminate fossil material. Shells are up to 4 mm in length and occur in great numbers in the specimens.

Remarks– These specimens are separated from the other bromalites based on an indefinite flattened form with a very thin groundmass, containing scattered inclusions that extend beyond the groundmass. On this basis they are tentatively identified as regurgitations (McAllister 1996). Identification of regurgitalites may be tested with geochemistry (Gordon *et al.* 2020; Robin *et al.* 2022).

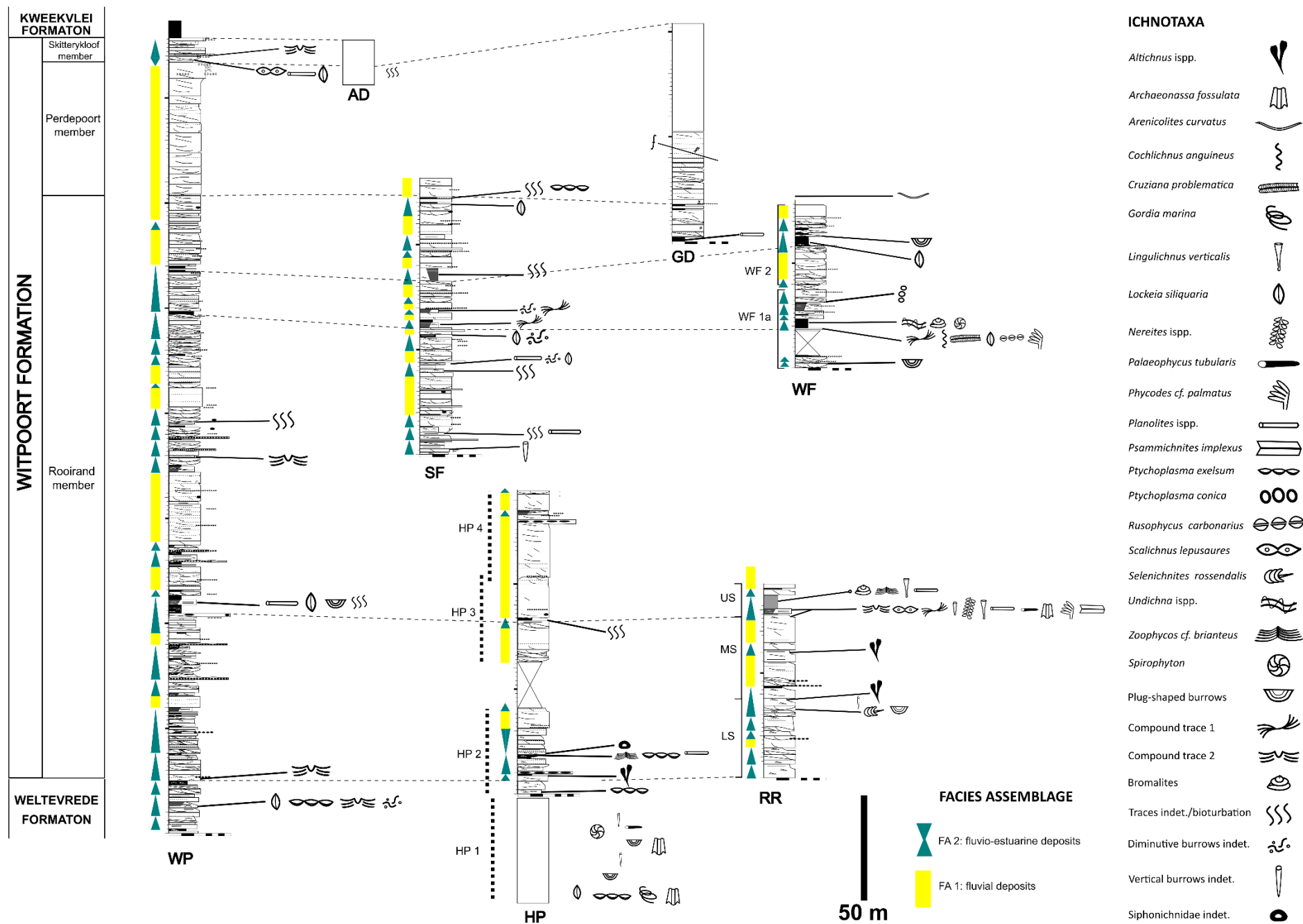


Figure 6.32: (previous page) Distribution of trace fossils in the study area**6.3. Discussion**

The described trace fossils give evidence for animal behaviours in the Witpoort Formation and the underlying upper Weltevrede Formation. They include evidence for activities of unspecified molluscs (*Archaeonassa*, *Psammichnites*), bivalves (*Lockeia*, *Ptychoplasma*, ‘compound traces 1 and 2’, *Scalichnus*; and indeterminate ‘Siphonichnidae’), arthropods (*Cruziana*, *Rusophycus* and *Selenichnites*), worms (*Zoophycos*, ‘*Spirophyton*’, *Nereites*, *Cochlichnus*, *Arenicolites* and *Altichnus*), lingulid brachiopods (*Lingulichnus*) and fish (*Undichna* and ‘bromalites’). The ichnoassemblages are dominated by shallow-tier burrows, whereas surface trails are rare. All the above are subsurface burrows with the exception of one invertebrate ichnogenus, *Archaeonassa*, and the vertebrate traces. This accords with typical biases in the sedimentary record, which favours the preservation of surface trails only under specific conditions (Mángano *et al.* 2024). True substrates or palaeosurfaces preserve under strict sets of conditions, favoured by sedimentary stasis (Davies and Shillito 2018), algal and mucous binding, absence of deeper tier bioturbators and bulldozers, high depositional rates, low hydrodynamic energy (with mud draping) and early carbonate cementation (Mángano *et al.* 2024). Mud-draped, rippled palaeosurfaces that appear to meet the criteria of true substrates (cf. Buatois *et al.* 2025) are common in the Witpoort Formation, but seldom preserve surface traces. It therefore appears that sporadic distributions of trace fossils and the paucity of surface traces are not taphonomic artefacts.

Among the inferred tracemakers throughout the stratigraphic column (Fig. 6.32) the most prolific were molluscs in both ichnodiversity and in the frequency of occurrence. The predominance of traces likely formed by molluscs is interesting as this group has a reputation for adaptation to stressful environments (Dalrymple and Choi 2007). Amongst shelly marine invertebrates, bivalves show the greatest tolerance of reduced salinities, with many species

experimentally shown to survive salinity levels of 5‰ or less (Castagna and Chanley 1973; Hammond 1983; Matthews and Fairweather 2004). The traces made by bivalves are in many cases distinctive, and have received much attention in recent years (López Cabrera *et al.* 2019; Buatois *et al.* 2017; la Croix *et al.* 2022). By analogue with the infaunal activities of recent bivalves much about bivalve ecology is understood by their ancient traces (Bradfield and Newell 1961; Stanley 1968; Hughes 1969; Kranz 1974; Wikander 1980; Bender and Davis 1984; Seilacher-Drexler and Seilacher 1999; Bromley *et al.* 2003). Extant bivalves provide reliable actuopalaeontological models because most higher taxa have extant representatives and their group life histories have remained relatively consistent since the Palaeozoic (Stanley 1968). Modern day marine bivalves are some of the more resilient and adaptable marine species, and are particularly successful in modern estuarine environments (Day *et al.* 2012). Some estuarine species exhibit tolerance of the various environmental stresses inherent to the environment, for example euryhalinity (Bertrand *et al.* 2017; Pourmozaffar *et al.* 2020) and escape potential, particularly in families that are adapted for deposit-feeding such as Tellenida and Protobranchia (Kranz 1974). Moreover, an ability to shift feeding strategies with changing conditions is particularly conducive to estuarine or intertidal conditions. Intertidal bivalve species *Macoma balthica* and *Scrobicularia plana* can switch between suspension-feeding and deposit-feeding according to the microenvironment or tidal regime (Bradfield and Newell 1961; Hughes 1969).

Bivalvian tracemakers are responsible for an increasing number of compound traces in the rock record, which give evidence for various activities by a tracemaker (Ekdale and Bromley 2001; Bromley *et al.* 2003; López Cabrera *et al.* 2023). Presumed bivalvian mobile deposit-feeding traces (compound traces 1 and 2) are among the more frequently occurring and abundant traces in the Witpoort Formation, and potentially represent bedding surface and profile views of the same trace respectively. They appear to represent a similar, but less

systematic set of behaviours to the tellenid trace fossil *Hillichnus* (Bromley *et al.* 2003). Diagnosing its full ramification, in light of what is known on modern and ancient bivalve behaviour, potentially can result in identification of a tracemaker at family or even species level, which for now remains uncertain. *Ptychoplasma* is an abundant trace at certain limited horizons in the upper Weltevrede Formation and Witpoort Formation. Co-occurrence of these locomotory traces with vertical siphon traces and an unidentified bivalve at Coombs Hill provides further potential for linking bivalvian compound traces with their tracemakers. Integration of bivalve taxonomy and ichnological diversity will give insights about their ecology and distribution in the Palaeozoic. For the present analysis, it is clear that characteristic traces of cleft-foot bivalves (*Protovirgularia*) unknown in the Witpoort Formation succession, whereas they are very common in Carboniferous and younger intertidal deposits (Mángano *et al.* 2002b).

In terms of ichnotaxonomic diversity, the Witpoort Formation has some overlap with the Carboniferous brackish water ichnofaunas of North America with ichnotaxa such as *Nerietes* spp. (Hakes 1976; Rindsberg 1994; Mángano *et al.* 2002b), *Cruziana problematica* (Mángano *et al.* 2002b), *Selenichnites rossendalensis* (King *et al.* 2019), *Lockeia siliquaria* (Rindsberg 1994) and *Psammichnites implexus* (Mángano *et al.* 2002a), all of which are common in Carboniferous estuarine deposits. They are recorded in low-diversity assemblages in the Witpoort Formation, normally dominated by a single shallow infaunal deposit-feeding ichnotaxon. Though the dominating ichnogenus is different at almost every studied stratigraphic level, collectively these ichnotaxa represent the depauperate *Cruziana* ichnofacies (Buatois *et al.* 2005). The paucity of suspension-feeding traces in proximal *Cruziana* ichnofacies systems such as estuaries is explained by elevated turbidity in the water column (Buatois *et al.* 1997b).

The ichnotaxonomic diversity and behavioural strategies are therefore consistent with opportunistic behaviour under stressful, tidally-influenced conditions. Organism diversity in such systems tends to be lowest close to the fluvial–tidal transition (Gingras *et al.* 1999; Dalrymple and Choi 2007; Diez-Canseco *et al.* 2015), with salinity levels at between 1–5‰, from where diversity increases seaward and landward. Bioturbation intensity is highly variable and sporadic in intertidal deposits, ranging from BI 0–2 (Gingras *et al.* 1999; Dalrymple and Choi 2007). This distinguishes brackish water deposits from open-marine deposits, where bioturbation is more regular and intense (Dalrymple and Choi 2007). Additionally, organisms in this environment tend to be reduced in size compared to marine counterparts (Buatois *et al.* 2005; Dalrymple and Choi 2007). These characteristics are the result of stresses inherent to the intertidal environment, including fluctuating salinity, regular exposure, elevated suspended sediment concentrations and rates of deposition (Diez Canseco *et al.* 2015; Gingras *et al.* 2016).

It is not possible with present evidence to make a detailed comparison between diversity gradients and trace fossil size in the Witpoort Formation with contemporary open-marine deposits from the sedimentary basin, which is the normal approach for diagnosing brackish water systems (MacEachern 1994; Buatois *et al.* 2005; MacEachern and Gingras 2007). It is, however, noted that the ichnotaxa recorded here are similarly sized to those found in Carboniferous intertidal deposits. These generalisations should be mediated with a more detailed look at specific trace fossil occurrences.

Three trace fossil-bearing horizons in the study area are studied based on museum collections derived from intensive palaeontological excavations, whereas the remainder were located at surface during fieldwork. All three of these studied horizons occupy heterolithic sandstone underlying prominent fossil-bearing shale or mudstone deposits. At each locality a different ichnogenus dominates; at Waterloo Farm, *Cruziana problematica* intergrading with

Rusophycus carbonarius; at Rabbit Ridge, *Nereites* ispp. and at Coombs Hill, *Ptychoplasma excelsum*.

A *Cruziana problematica*–*Rusophycus carbonarius* association is exclusively recognised in the Waterloo Farm section (WF 1a) in the study area. Trace fossil-bearing slabs were collected from a < one metre-thick unit of thin-bedded, undulatory micaceous sandstone interbedded with thick reddish mud drapes, which closely underlies black shale at Waterloo Farm (R.W. Gess pers. comm.). This interval is not presently exposed, but was studied by Hiller and Taylor (1992, Fig. 11) who interpreted the thin-bedded sandstone unit as a washover deposit. In the present analysis, based on the studied exposures, and by comparison to similar facies elsewhere in the study area (FA 2b in Chapter 5.4.2), the deposit is interpreted as estuarine flats within a fluvial–tidal transition zone. The rhythmic sandstone–mudstone motif potentially represents a fluvial flood-dominated setting with estuarine mud settling during the interflood periods (e.g. Van den Berg *et al.* 2007; Jablonski and Dalrymple 2016).

Cruziana and *Rusophycus* predominate on multiple bedding surfaces, usually in sub-parallel orientation. Their direction of movement is unknown as the slabs were studied *ex situ*. Uncommon traces from the same beds include small *Lockeia*, *Cochlichnus*, *Phycodes*, ‘plug-shaped burrows’, and ‘compound trace 1’. Bioturbation is almost exclusively bedding-parallel (BI 0, BPBI 1–3). This assemblage does not have an unequivocal brackish water signature, as it lacks typical marine traces, but contains ichnotaxa that are well represented in Palaeozoic non-marine deposits (i.e. *Cochlichnus anguineus* and *Cruziana problematica*). These are facies crossing traces, which are also known from intertidal and marine deposits. They potentially represent a freshwater inner estuary, or, possibly, the onset of brackish conditions.

The overlying estuary-basin black shale contains abundant evidence for brackish water, including monospecific, densely bioturbated ‘*Spirophyton*’ beds, which are typical of Devonian marginal-marine assemblages (Miller and Johnson 1981; Miller 1991; Uchman *et al.* 2004) and an estuarine flora and fauna (Gess and Whitfield 2020).

The Rabbit Ridge ichnofauna occupies heterolithic sandstone facies underlying and interbedded with a ~10 m thick mudstone unit (RR upper section). By contrast with the Waterloo Farm deposits, it contains clearer evidence for elevated salinity. Thin-bedded, parallel-laminated sandstone with mud drapes underlie and are interbedded with the mudstone layer. Bioturbation is absent to moderately profuse on bedding surfaces (BPBI 1–3), whereas vertical burrows occur sporadically, in low density (BI 0–1). Certain lower and upper bounding surfaces of sandstone beds are dominated by *Nereites* spp., representing a range of ichnospecies that intergrade. They occur with *Planolites* and ‘compound traces 1 and 2’. Uncommon ichnotaxa include *Archaeonassa fossulata* and *Psammichnites implexus*, cf. *Scalichnus lepusaures*, *Lingulichnus* and *Palaeophycus*. The low ichnotaxonomic diversity (1–3 ichnogenera per bedding plane), sporadic distribution of traces and dominance of subsurface deposit-feeding structures with a small proportion of vertical burrows point to a depauperate *Cruziana*–*Skolithos* ichnofacies assemblage, consistent with a brackish setting (Diez-Canseco *et al.* 2015; Shchepetkina *et al.* 2019). *Zoophycos* and *Lingulichnus* occur in the overlying mudstone, associated with a monospecific lingulid brachiopod community.

Similarly, at Coombs Hill, trace fossil concentrations are found in heterolithic facies and overlying mudstone, but are uncommon in the remainder of the succession. The 3 m thick Inner Outcrop mudstone horizon is underlain by a laterally variable heterolithic unit. Medium-bedded sandstone with rhythmic parallel-lamination, gutter casts and rhythmic mud-drapes and double drapes is locally developed within a metre-thick unit underlying the mudstone. This facies is burrowed with a low intensity of vertical structures (BI 0–1),

including *Scalichnus lepusaures* and possibly associated *equilibrichnia* ('plug-shaped burrows'). The upper 30 cm of the deposit comprises thin-bedded, weakly-laminated sandstones with muddy partings, in which indeterminate vertical to inclined burrows are concentrated in discreet intervals of 5–10 cm and sandstone–mudstone contacts are dominated by *Ptychoplasma exelsum* (hypichnial) and *Archaeonassa fossulata* (epichnial). Articulated bivalve shells are common in this unit and in the overlying mudstone in which, on certain bedding surfaces, they occur in great numbers. The trace and body fossils indicate a bivalve-dominated intertidal assemblage at this horizon.

The three described stratigraphic levels all have different ichnotaxonomic signatures, yet the commonalities are evident. The heterolithic sandstones that underlie mudstone and shale facies at these sites contain intertidal trace fossil assemblages, with ichnotaxonomic variations that probably reflect a spectrum of salinity levels, depositional rates and suspended sediment concentrations within fluvial–tidal transition zones. Paradoxically, fluvially-generated sandstones in this type of environment may be intruded by marine trace fossils due to the upstream shifting of the tidal maximum during periods of low river discharge (Dalrymple *et al.* 2015). Brackish water conditions may account for the localised, low-diversity occurrences of trace fossils in heterolithic and mud-rich intervals throughout the Witpoort Formation stratigraphic column. Their apparent absence in sandstone-dominated successions is indicative of conditions that were inconducive to trace-fossil preservation. The highly amalgamated sheet-like strata of FA 1 were generated in braided river rather than open-marine environments, explaining the total absence of marine invertebrate body and trace fossils in this facies assemblage. Invertebrate fauna may have been subdued by harsh sub-polar conditions, high-energy flashy discharge and the absence of sheltered overbank environments, however, if non-marine invertebrates were active in this fluvial setting, their traces may have had a low-preservation potential due to the intense amalgamation of strata.

Ichnotaxonomy may help to resolve the palaeoenvironments of the underlying Weltevrede Formation, which is alternatively interpreted as open-marine (Cole and Labuschagne 1983; Cotter 2000), or deltaic to intertidal (Hiller 1990). Trace fossils in the uppermost strata of this formation show a great ichnotaxonomic overlap with the Witpoort Formation. They include *Ptychoplasma excelsum*, *Lockeia siliquaria*, ‘*Spirophyton*’, ‘plug-shaped burrows’, ‘compound trace 2’, *Palaeophycus*, and *Gordia marina*. Only the last of them is unknown in the Witpoort Formation. This cursory analysis would favour palaeoenvironmental congruence between the limited heterolithic horizons of the Witpoort Formation and the far more elaborately developed Weltevrede Formation’s rhythmites, which, as already noted by Johnson (1976), share a strong resemblance.

**CHAPTER 7 – INSIGHTS FROM A MONOSPECIFIC LINGULID BRACHIOPOD BED IN THE
LATE DEVONIAN OF SOUTH AFRICA**

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Author contributions:

Harris – Conducted the research, wrote the manuscript

Gess – Supervisor, original conceptualisation, discovery of the locality, reviewed and edited the manuscript

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7.1. Abstract

A formerly unknown fossil bearing locality in the lower part of the Witpoort Formation (Witteberg Group, Cape Supergroup) is described from the Eastern Cape (South Africa). Uniquely for these strata, it provides evidence for a back-barrier lagoon hosting a monospecific lingulid brachiopod fauna. This represents the youngest record of marine invertebrates within the Cape Supergroup. The occurrence extends the age range of the genus *Dignomia* Hall into the Famennian Stage, contrary to previous estimates of an Ordovician to mid-Devonian range. Abundant lingulid brachiopod valves are associated with plant and fish remains within a laterally extensive, up to 9-metre-thick mudstone. Shell compressions are concentrated in patches within the upper portion of the horizon and are preserved parallel to bedding as well as perpendicular and oblique to bedding (apparently *in situ*). Plant fossils exhibit a greater degree of transport than in other studied Witpoort Formation mudrock layers (Coombs Hill and Waterloo Farm). The monospecificity of the shelly invertebrate fauna derives from environmental stress and lingulid brachiopods' unusual ability to tolerate

fluctuating environmental conditions, including salinity, oxygen levels, and temperature.

Ichnofauna within the deposit include, among others, *Nereites* and '*Spirophyton*', suggesting opportunistic colonisation of the substrate by deposit-feeding invertebrates in a marine-dominated backshore setting. The Late Devonian was a time of global environmental disequilibrium, characterised by periodic flooding of continental margins and massive biotic overturn. Ongoing research into eustatic and environmental changes recorded within the Witpoort Formation uniquely provide insights into high-latitude conditions during the Famennian.

7.2. Introduction

Lingulid brachiopods have an almost unparalleled fossil history, extending over 500 million years from the early Cambrian to the present (Yugan *et al.* 1993; Emig 1997). This group is an extreme example of phenotypic conservatism with numerous studies indicating continuity of lifestyle and physiology between Palaeozoic and Recent taxa (Craig 1952; Ferguson 1963; Paine 1970; Savazzi 1986; Zonneveld *et al.* 2007; Zabini *et al.* 2012). Lingulidae (the only infaunal brachiopod taxon) characteristically live within mucus-lined burrows ranging from the intertidal zone to neritic substrates (Savazzi 1986; Kowalewski 1996; Emig 1997; Zonneveld *et al.* 2007). Their palaeoenvironmental context is informed by research on the ecology of extant lingulids, which exhibit profound resilience to environmental fluctuations (Morse 1902; Craig 1952; Chuang 1961; Paine 1970; Hammen and Lum 1977; Kenchington and Hammond 1978; Hammond 1983; Kowalewski 1996; Emig 1997). Stressful, marginal environmental conditions and extinction event aftermaths therefore facilitate superabundant lingulid occurrences (Rodland and Bottjer 2001; Zonneveld *et al.* 2007).

Palaeoecological study of Late Devonian sequences contributes to understanding of significant evolutionary developments, climatic changes and associated extinction events that

characterise this epoch (Algeo and Scheckler 1998; Bond and Wignall 2008; Carmichael *et al.* 2019; Percival *et al.* 2019). As yet, most Late Devonian research has focussed on strata of formerly low-latitude Laurussian provenance. Southwestern Gondwana (which included the southern portions of Africa and South America) was, by contrast, situated in a sub-polar setting. As most South American and Falkland Islands successions of this age were regionally truncated by a massive Permo-Carboniferous glacial event (de Vargas *et al.* 2020), the South African Witteberg Group almost uniquely retains a sub-polar succession spanning this interval.

The Famennian aged Witpoort Formation (of the middle Witteberg Group) records a globally correlatable regressive eustatic signal, locally characterised by shallowing and shoreline progradation (Cooper 1986; Hiller and Taylor 1992). The predominantly quartzitic strata of this formation contain pelitic interbeds preserving unique evidence regarding high-latitude marginal-marine ecosystems. Best studied of these is the Waterloo Farm lagerstätte, an estuarine, tetrapod-bearing locality exposed by roadworks in 1985 (Gess and Whitfield, 2020). Comparable ecosystems are evidenced in strata of similar facies, revealed by roadworks in 2015 (Harris *et al.* 2021). Here we present evidence for an autochthonous, monospecific lingulid brachiopod fauna, recovered from a mudstone horizon extending over several kilometres.

7.3. Geological setting

The Witpoort Formation is the middle unit of the Witteberg Group, which comprises the uppermost subdivision of the Cape Supergroup. Deposition of the Cape Supergroup was initiated by opening of the Agulhas Basin in the Cambrian and was near continuous until the mid-Carboniferous. The entire package was uplifted during the late Palaeozoic by northward-directed compressional tectonics following subduction to the south (Shone and Booth 2005).

In the absence of high resolution biostratigraphic indicators, eustatic shifts have provided the most accurate framework for Cape Supergroup geochronology (Cooper 1986) and the texturally mature sand shoals of the Witpoort Formation are determined, according to event stratigraphy, to coincide with the Famennian regressive interval (Cooper 1986; Hiller and Taylor 1992). A Famennian age for the Witpoort Formation is supported by vertebrate biostratigraphy (Gess 2016) as well as Lower Carboniferous palynomorphs from the overlying Lake Mentz Subgroup (Streel and Theron 1999). The Witpoort Formation comprises a 300–350 m thick package of mature quartzite with minor interbedded mudrock (Hiller and Taylor 1992). Though earlier research has indicated localised thicknesses of 800 metres, or more, in the region of Makhanda (Thamm and Johnson 2006), apparent thickening of stratigraphic units eastwards in the Cape Fold Belt is likely related to structural duplication (Shone and Booth 2005). The Witpoort Formation is currently considered to represent deposition along an east-west trending, wave-dominated coastline (Hiller and Taylor 1992; Thamm and Johnson 2006) with beach-barrier complex deposits comprising shoreface, beach, tidal inlet channel, washover fan and lagoonal sediments (Hiller and Taylor 1992). Its lower contact is established as the base of the first prominent mature arenitic stratum overlying the predominantly argillaceous upper Weltevrede Formation (Johnson *et al.* 2006).

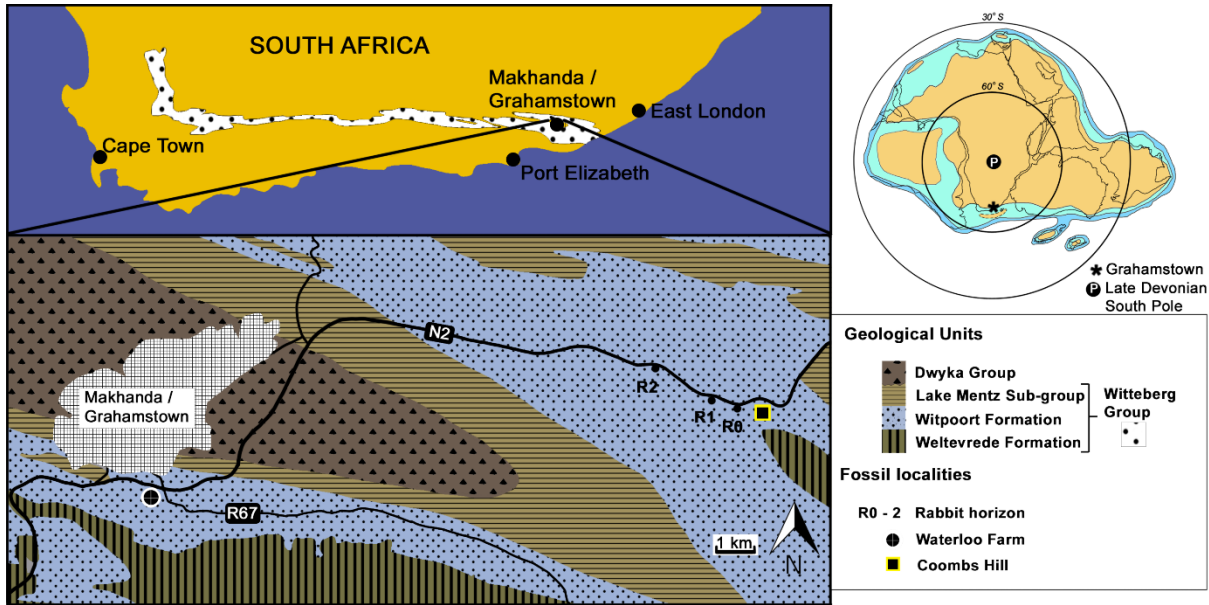


Figure 7.1: Location of study area. Geological map (below) of the Makhanda/Grahamstown region (based on Council for Geoscience, sheet 3326 (1: 250 000)), showing the locations of lingulid bearing outcrops, R0, R1 and R2, and fossil localities at Waterloo Farm and Coombs Hill. Stratigraphic units and their ages are provided at bottom right. Palaeogeographic reconstruction (top right) modified from Gess and Ahlberg (2018).

7.4. Material and methods

A new fossil locality in the Witpoort Formation was located along the ridge known as Green Hills (informally Rabbit Ridge) during upgrades to the N2 highway in 2015. It comprises an unusually thick carbonaceous mudstone horizon, up to 9 or more metres thick, exposed in steeply tilted strata beneath the roadbed. This 'Rabbit horizon' was traced for 4.6 km along strike, along which it exhibits variable thickness between localities R0, R1 and R2 (Fig. 7.1). Stratigraphic logging of exposed strata on this ridge was conducted to establish geological context and depositional setting of the fossiliferous mudstone horizon.

Stratigraphic correlation between outcrop R0 (33°17'57.8"S 26°45'2.2"E) and R1 (33°17'47.6"S 26°44'29.5"E) was made by direct observation during undercuttings to the road bed. R2 (33°17'18"S 26°43'25.6"E) was correlated along structural strike, and is interpreted as forming part of the same horizon. The N2 roadway follows the layer for a

measured distance of 700 m (from R0 to R1, Fig. 7.1), whereafter it runs parallel to the layer, before crossing it again approximately 3 km westward of R0, at outcrop R2.

At R0 this argillaceous horizon is 2.5 m thick and it thickens, over 700 m along strike, to 9 m at R1 (Fig. 7.4C, D) where a stratigraphic log of the 'upper section' was obtained (Fig. 7.2). At R2 the mudstone horizon is at least 8 m thick. Lingulid brachiopod fossils occur at outcrops R0, R1 and R2 dispersed over 3.5 km. At both R1 and R2, concentration zones of lingulid shells similarly occur near the top of the Rabbit horizon, and a single lingulid shell was found within spoil derived from R0. Further cuttings into this mudstone horizon around 1 km west of R2 are apparently unfossiliferous.

Fossil material was systematically collected by the authors from three localities (R0, R1 and R2) when abundant fresh rock was exposed during road excavations between November 2015 and September 2016. Specimens were recovered at outcrop and to a lesser extent from 'spoil' removed from the layer during roadworks. All collected material is housed at the Albany Museum (Earth Sciences Department - Devonian Lab) in Makhanda (formerly Grahamstown).

Collected fossil lingulid valves were selected for morphometric analysis where bedding parallel, disarticulated and whole. The lengths and widths of valves were measured to 0.1 mm certainty using a dial calliper. This data is recorded in Supplementary Material A (Appendix 4A). Length was taken along the midline running from the umbo to the anterior margin of the valve. Width was measured at the widest point of the shell perpendicular to the bilateral midline.

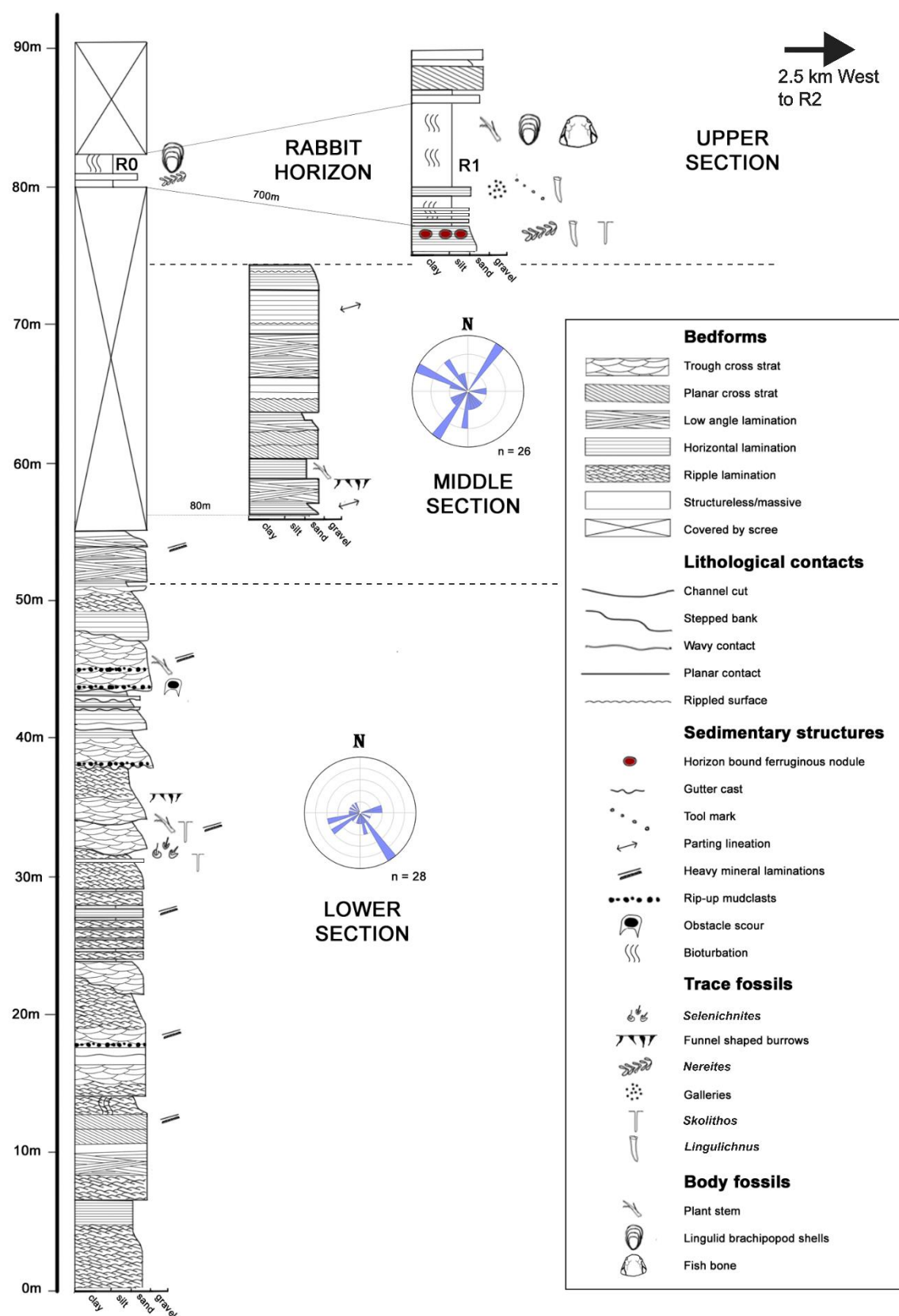


Figure 7.2: Interpretive stratigraphic log diagram of the rock outcrops exposed on Rabbit Ridge.

7.5. Results

7.5.1. Geology

Geological Overview

The Rabbit horizon is interbedded between thick arenaceous stratigraphic packages. Evidenced in road cuttings are eighty stratigraphic metres of predominantly arenaceous strata underlying the Rabbit horizon, which represent the basal portion of the Witpoort Formation. Strata exhibit upward younging but are structurally tilted, dipping between 50° and 80° to the southwest, with a strike of between 110° and 130°. Rock exposures along this ridge include old and new road cuttings, providing a near complete 90 m thick lithological section, from which an interpretive log is derived (Fig. 7.2). The succession was divided into three sections (lower, middle and upper) based on lithological character.

The lower section begins at, or near, the mappable base of the Witpoort Formation, corresponding to the first thick, mature quartzitic stratum overlying the Weltevrede Formation. A complete lithological section from the base of the lower section to the top of the upper section at R0 was exposed at the beginning of our study (November, 2015) in an old road cutting. Expansion of the road, and partial destruction of the old cutting, has obscured the middle and upper part of the outcrop with scree; however, correlative sections from the opposite side of the road (middle section) and further along the extensive mudstone horizon (upper section) fill in much of the concealed lithology.

Lower Section

Lithology– This section comprises quartz arenite, greywacke and siltstone forming several erosively based 3–10 m thick upward-fining units (Fig. 7.3A). Thickness and maturity of sandstone beds decrease upwards within each unit, whereas interbedded siltstone layers increase in thickness. Sandstones are generally texturally mature, structureless to planar and trough cross-stratified, developed in tabular, lenticular or wedge-shaped sets, commonly

around 30 cm, and up to 1 m thick. Ripple cross-lamination (including sub-critical climbing ripples, Fig. 7.3B) and horizontal lamination are developed in sandstone towards the upper portions of cycles, which commonly have thin structureless siltstone interbeds.

Heavy mineral laminations and reactivation surfaces commonly occur in sandstone beds, and occasional thin (mm scale) clay drapes are preserved on cross-bed foresets. At the base of one thick trough cross-stratified unit, mudstone intraclasts, of up to cobble size, are surrounded by obstacle scours (Fig. 7.3C) and current drag tails. Intraformational mudclast conglomerate and plant stems (some identifiable as lycopods, Fig. 7.3G) are common at or near the erosive bases of units. Stepped erosional surfaces, with downcutting relief of up to 1 m are observed at two horizons and shallow erosional scours are visible at several horizons (Fig. 7.2). The lower section terminates where ripple and trough cross-stratification become scarce and are succeeded by tabular strata of the middle section at around metre 50 in the log (Fig. 7.2).

Ichnology– Evidence for bioturbation is rare in the lower section, and trace fossils are limited to a few monospecific horizons. Numerous inclined to vertical, pipette-shaped burrows occur within one ripple cross-laminated horizon. Burrows are 10–25 mm in maximum diameter, tapering downwards to a narrow, J-shaped or Y-shaped (upwards branching) basal tube. An erosion surface truncates many of these burrows, which have subsequently been re-established at a slightly higher horizon (Fig. 7.3E).

Similar traces from the Lower Devonian of Bolivia assigned to *Altichnus* spp. were interpreted to indicate lower intertidal environments (Gaillard and Racheboeuf 2006). The burrows resemble *Altichnus* in the apparent upward branching of the basal tube and in exhibiting re-equilibrium behaviours (Bromley and Hanken 1991; Gaillard and Racheboeuf 2006). They are clearly distinguished from other funnel- or pipette-shaped burrows, *Lingulichnus* and *Rosselia*, in being unlined and lacking complex concentric infill (Hakes

1976; Uchman and Krenmeyer 1995; Zonneveld *et al.* 2007). *Altichnus* has been compared to *Monocraterion*, and transitional forms with *Altichnus* have been recorded (Gaillard and Racheboeuf 2006), but simple funnel-shaped burrows are excluded from *Monocraterion* as the type material shows radiating tubular structures emanating from the burrows (Jensen 1997).

A thick trough cross-bedded sandstone unit towards the top of the lower section preserves numerous hypichnial burrow casts where it overlies siltstone of the previous cycle. Each burrow is horseshoe-shaped (around 1 cm in width), with some preserving a median posterior ridge (Fig. 7.3D, F). These occur in serial alignment or are isolated. In both the shape and occurrence in serial alignment, the traces resemble *Selenichnites* (Morrissey and Braddy 2004; Romano and Whyte 2015). All traces are orientated similarly, with the presumed anterior margin facing a general south westerly direction. *Selenichnites* has been recorded in fluvial (Morrissey and Braddy 2004), intertidal (Morrissey *et al.* 2012), and fully marine strata (Romano and Whyte 2015).

Interpretation– The lower section is interpreted as largely comprising deposits of channel cut, fill and abandonment, as evidenced by upward-fining cyclical units having scoured bases with lag deposits, predominant trough cross-stratification and lenticular geometry of sandstones (Miall 2014). This is consistent with previous facies interpretations for the lower part of the Witpoort Formation where laterally migrating, ebb-dominated tidal channels are considered to have reworked much of the depositional record of foreshore and shoreface environments (Hiller and Taylor 1992).

Palaeocurrent indicators from cross-strata in the lower section are unimodal (Fig. 7.2), exhibiting a wide splay, predominantly directed towards the south. Assuming an east-west trending shoreline, the direction of sediment transport is interpreted as seawards.

Evidence for a tidal channel origin for the lower section is thus equivocal, given the absence of clearly bimodal current indicators or tidal bundles and the paucity of trace fossils (Visser 1980; Thomas *et al.* 1987; Cheel and Leckie 1990; Dalrymple 2010). Though rare monospecific trace fossil horizons (*Altichnus* and *Selenichnites*) are consistent with a brackish water setting, aggradational stacking of 50 metres of tidal channel deposits would be uncommon. An alternate interpretation, consistent with evidence from the lower section, is deposition on a coastal braidplain influenced by episodes of marine transgression.

Middle Section

Lithology– The 25-metre-thick, lithologically uniform middle section is characterised by tabular, laterally persistent planar cross-stratified and horizontally laminated sandstone (Fig. 7.4A). Parting lineations are evident on bedding planes of more horizontally laminated strata. Ripple cross-lamination is scarce in this section, yet unidirectional, sinuous ripple form crests do occur on the upper bounding surfaces of sandstone sets (Fig. 7.4B). Large erosive troughs and hummocks with wavelengths up to 3.5 m truncate a few horizons, producing wavy set surfaces. Convolute bedding is observed at one horizon. Upward-fining sandstone-siltstone cycles that characterise the lower section are absent from the middle section. Argillaceous sediments are confined to two thin, horizontally laminated siltstone horizons lower in the section, one of which preserves comminuted plant material.

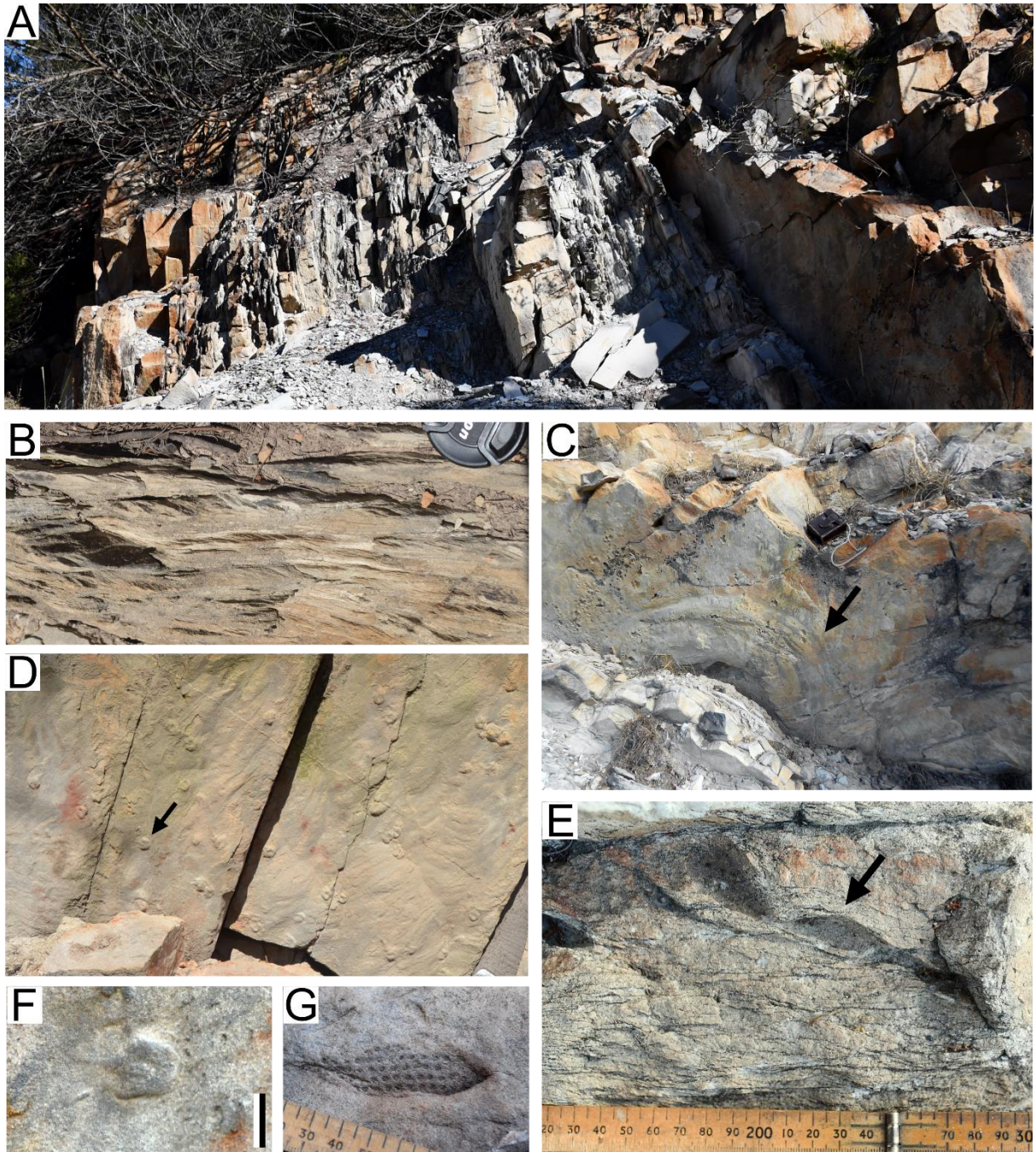


Figure 7.3: Geology of the lower section on Rabbit Ridge: A) A roughly 3m thick fining upward unit, (younging direction is to the right). B) Ripple laminated fine-grained sandstone showing a low angle of ripple-climb. Camera lens cover is 55 mm in diameter. C) The erosive base of a unit showing weathered out mudclasts and associated obstacle scours. Arrow indicates a large scour surrounding a cobble-sized mudclast. Geological compass for scale. D) Numerous hypichnial resting traces (cf. *Selenichnites*). Arrow indicates the trace enlarged in F. E) Pipette-shaped burrows (cf. *Altichnus*) exhibiting erosive truncation (arrow) and re-establishment at a higher horizon (scale in cm). F) Enlargement of a single resting trace from D. Scale bar = 1 cm. G) A fragment of plant axis (lycopod) in sandstone of the lower section.



Figure 7.4: Geology of the middle and upper sections on Rabbit Ridge: A) Planar cross-stratified and horizontally laminated sandstone from the middle section. Geological hammer for scale. B) Sinuous current ripple form sets from the middle section. Geological compass for scale. C) Demolition of the several metres thick Rabbit mudstone at R1. Arrow indicates an orange oxide-rich marker horizon in the lower-middle part of the layer. Excavator machine in the background provides a sense of scale. D) One of the authors excavating lingulid brachiopods from the upper part of the Rabbit horizon mudstone. Note that the former roadway overlies the mudstone layer.

Ichnology– Simple vertical, funnel-shaped burrows are exposed on a planar bedding surface underlying horizontally-laminated siltstone near the base of the middle section. Around 50 burrows occur in a square metre of bedding surface. These are consistent with a plan view of burrows compared to *Altichnus* from the lower section (see above). Burrows in both horizons are preserved as hollowed-out moulds, are similar in diameter, are erosively truncated, and occur in a similar density, suggesting palaeoecological continuity from the lower to middle sections. Trace fossils were not observed in the remainder of the middle section.

Interpretation– The middle section is characterised by a tabular, sheet-like sandstone geometry, reminiscent of much of the depositional record of the Witpoort Formation (Hiller and Taylor 1992; Cotter 2000). Existing palaeoenvironmental interpretations ascribe a shallow-marine origin to the stacked quartz arenites of this formation (Hiller and Taylor 1992; Cotter 2000; Thamm and Johnson 2006), although Broquet (1992) cautioned against use of sediment maturity to infer a shallow-marine origin for quartzites in the Cape Supergroup in the absence of clear marine indicators.

Although stacked sheetlike sandstones can occur within prograding shoreface deposits (e.g. Clifton 2006), they may also form in braided fluvial settings (Davies and Gibling 2010; Miall 2010). The stacked sandstone beds exhibit no upward coarsening trend or facies progressions indicative of upward shallowing in a prograding shoreface setting. Nor is hummocky/swaley cross-stratification expressed, even though erosive features superficially resembling hummocks and swales do occur. Aside from a localised occurrence of funnel shaped burrows (cf. *Altichnus*) at a sandstone–siltstone contact lower in the section, the only evidence favouring a marine setting is the polymodal orientation of planar foresets (albeit from a small data set). This, however, does not exclude fluvial settings since upper flow regime conditions may deposit upstream migrating foresets (Bridge and Demicco 2008).

No clearly marine indicating trace or invertebrate fossils are known from sheet-sandstone facies throughout the formation and this is a reasonable argument against a shoreface depositional setting (Broquet 1992). If a shallow-marine interpretation were correct, then the paucity of marine fossils and trace fossils in this type of succession can be explained by a high rate of deposition, erosion and reworking that exceeds the time required for invertebrate colonisation of the substrate (Hiller 1990). A fluvial setting is, however, considered more plausible for the sheet-sandstone facies. Minor interbedded siltstones representing lower energy episodes, associated with possible marine traces, indicate increased rate of accommodation, possibly during brief transgressions.

Upper Section

Lithology– The upper section comprises an upward-fining sequence. Sandstone (now deeply weathered) underlies the mudstone horizon at R1, grading upwards into gritty mudrock. Ferruginous concretions populate the base of the mudrock horizon. Throughout the lower half of the mudstone layer, subordinate interbeds of fine-grained sandstone range from a few millimetres up to decimetres thick. These interbeds display horizontal laminae to low angled and wave-ripple cross-laminae and preserve numerous trace fossils (see below). In cross-section, ripples exhibit bidirectional laminae and lateral pinching and swelling. One bedding plane preserves ripples with flattened crests, on which tiny interference ripples and finely preserved trace fossils are superimposed (Fig. 7.5D, G). Preservation of these structures may have been influenced by microbial binding of the sediment. A thicker sandy interbed is rich in oxidised iron, producing an orange marker horizon visible for tens of metres along strike (Fig. 7.4C). This horizontally-laminated stratum preserves tool marks but very few traces.

The upper part of the mudstone horizon comprises massive monolithic grey mudstone. The mudstone layer is truncated by a metre-thick medium grained massive sandstone unit,

fining upward to a thin (30 cm) muddy stratum. Overlying this, a thickly bedded planar cross-stratified mature blue-grey quartz arenite unit represents termination of argillaceous deposition. Several tens of metres of relatively homogenous tabular arenitic sandstone (which were not systematically logged) concludes the succession exposed on this ridge.

Ichnology– The upper section preserves a diversity of classically marine ichnofossils. Abundant well-preserved traces occur in thin layers of fine-grained sandstone interbedded within the lower half of the mudstone. These display little bioturbation and preserve thin laminae, despite bedding surfaces being dominated by abundant horizontal burrows (Fig. 7.5D). The ichnoassemblage of the sandy interbeds comprises a low diversity of non-diminutive (moderately sized) traces with moderate disparity. Bedding surfaces are dominated by one or a few ichnotaxa, including horizontal meniscate structures with lateral ridges, representing surface locomotion trails (*Archaeonassa*; Fig. 7.5D), infaunal grazing traces (cf. *Nereites*), simple horizontal (*Planolites*) and vertical burrows (cf. *Skolithos* and *Lingulichnus*), and diminutive, complex branching traces that are developed in discrete clusters.

Notably, certain horizons are dominated by paschichnia provisionally attributed to *Nereites* spp. These occur at the sandstone–mudstone interface in positive epichinial or hypichnial relief. They comprise pustules of reworked sediment in two regular or irregular rows with or without a median furrow (Fig. 7.5C).

Endichnia include various vertical to bedding-oblique simple burrows. Some of these resemble *Lingulichnus* in having a fusiform profile where intersecting a bedding plane, and a thick outer wall surrounding an internal cylinder of oxidised sediment (Hakes 1976; Buatois *et al.* 2005) (Fig. 7.5A, B).

Large-sized ‘*Spirophyton*’ (up to 148 mm in diameter) preserved in mudstone (Fig. 7.5E), were recovered from spoil dug from R1. This occurrence is interesting because although

‘*Spirophyton*’ is common in the Lower Witteberg Group (Theron 1960; Tankard and Barwis 1982; Hiller 1990; Cotter 2000), it has been considered rare in the Witpoort Formation (Theron 1960). The Waterloo Farm black shale from the upper part of the formation, however, preserves numerous generally diminutive ‘*Spirophyton*’ within a brackish water setting (Gess and Hiller 1995a).

The taxonomy of *Spirophyton* and *Zoophycos* is problematic. Helicoidal *Zoophycos*-like spreite structures are commonly included in *Spirophyton* (e.g. Miller 1991; Uchman *et al.* 2004), but are mainly distinguished on the basis of size (Miller 1991). Seilacher (2007) considered *Spirophyton* to be a junior synonym of *Zoophycos*, and this is supported by recent study of the *Zoophycos* type material (Olivero 2007). Although there is no clear basis for separation of these ichnogenera, we retain the junior name for clarity (pending their description) because it has been used to identify traces of similar morphology in several studies on the Witteberg Group.

Trace fossils appear to have had a low preservation potential within the poorly bedded mudstone towards the top of the Rabbit horizon. Trace fossils include possible *Lingulichnus*, simple horizontal burrows, as well as spiral-shaped traces (Fig. 7.5F) resembling possible fish coprolites from Waterloo Farm (Gess and Hiller 1995a). In no instance were burrow casts discerned around vertically-orientated lingulid shells (Fig. 7.6B, C). This lack of differentiation of the burrows probably reflects their passive fill being identical to the host sediment.

Interpretation– The upper section represents a quiet water depositional environment extending for more than 4 km parallel to the inferred palaeo-shoreline. It was characterised by suspension settling of mud with episodic introduction of fine sand. Distinguishing between backshore and offshore muds relies on sedimentary structures, ichnology and fossil content (Tankard and Barwis 1982; Bann *et al.* 2004). A backshore environmental

interpretation is favoured by low degrees of bioturbation, absence of hummocky cross-stratification or marine restricted invertebrate fossils and conformable upward-fining from underlying arenitic sandstones.

The variable thickness, lateral extent and textural homogeneity (at least in its upper part) of the Rabbit mudstone horizon suggest origination within an expansive lagoon, formed by extended flooding of an area landward of a beach barrier. This afforded protection from marine wave and current energy.

Thin sandy interbeds burrowed by marine infauna in the lower part of the mudstone unit indicate higher energy episodes, perhaps as a result of marine incursions on the backshore. Horizontally-laminated sand may indicate upper flow regime conditions (Reineck and Singh 1980; Collinson *et al.* 2006). Ripple laminae observed in cross section exhibit features that are consistent with wave ripples (De Raaf *et al.* 1977).

Paucity of heterolithic facies in the Rabbit horizon in favour of monolithic mud deposition is consistent with very subdued tidal effects. Sandstone interbeds lower in the layer may have undergone periodic exposure, evidenced by flattening of ripple crests, horizon-bound ferruginous concretions and oxidation of sandy horizons. However, the lingulid-bearing monolithic mudstone exhibits none of these features, suggesting subaqueous deposition.

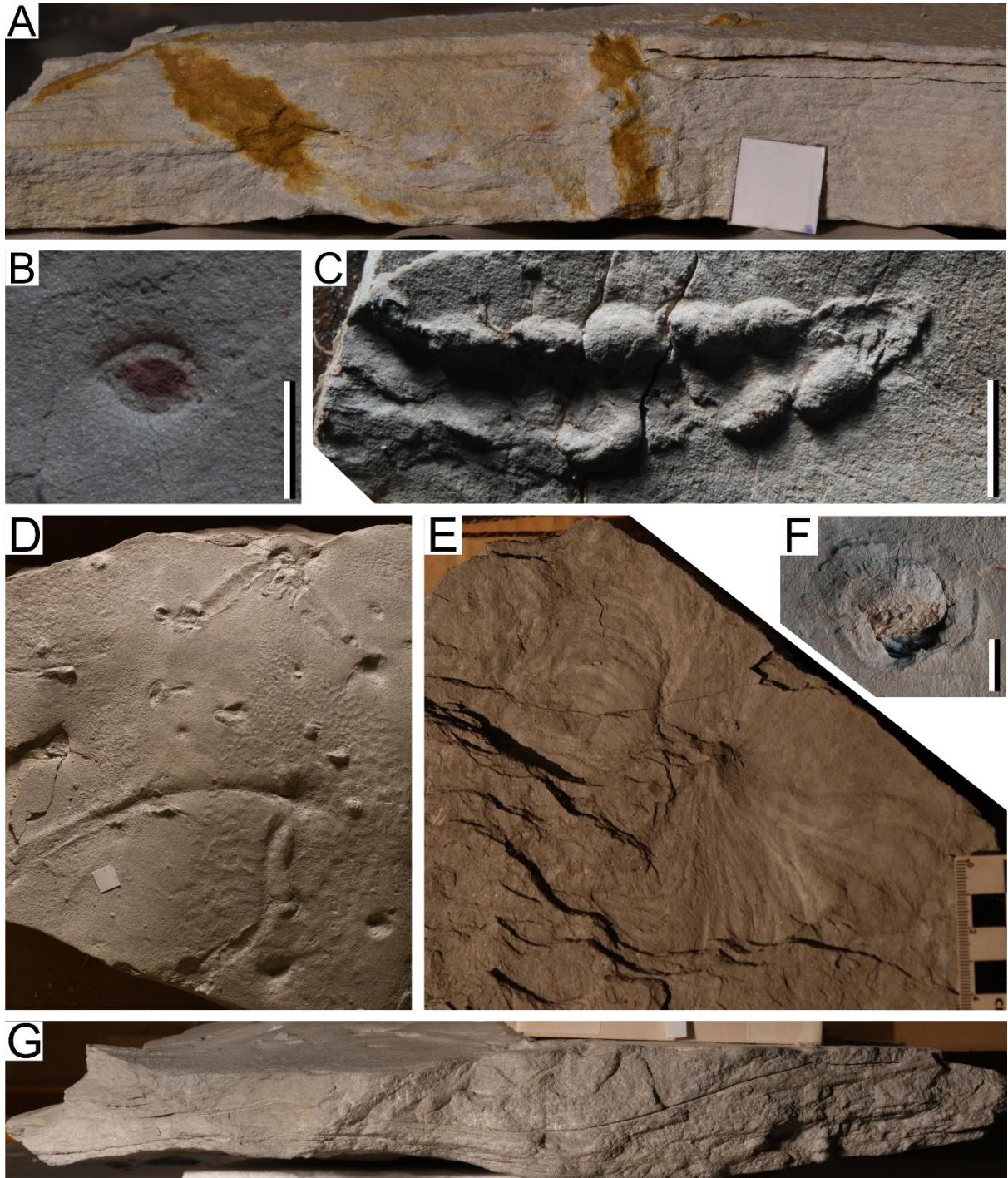


Figure 7.5: Trace fossils from the upper section on Rabbit Ridge: A) Vertical to inclined burrows (cf. *Lingulichnus*, AM 7786a) exposed in profile (scale square = 1 cm). B) Cross section of a vertical burrow (cf. *Lingulichnus*, AM 7791a), showing a thick outer wall and internal oxidation. C) Epichnial grazing trace (*Nereites*, AM 7790, scale bar = 1 cm). D) Rippled palaeosurface, burrowed by horizontal meniscate trails and vertical burrows, with small ladderback ripples formed upon the crests of larger flat-topped ripples (AM 7789, scale square at bottom-left is one cm). E) ‘*Spirophyton*’ in mudstone (AM 7783a, scale in cm). F) Spiral-shaped granular trace (possible fish coprolite, AM 7807, scale bar = 1 cm). G) Slab from (D) showing flat-crested ripples in profile.

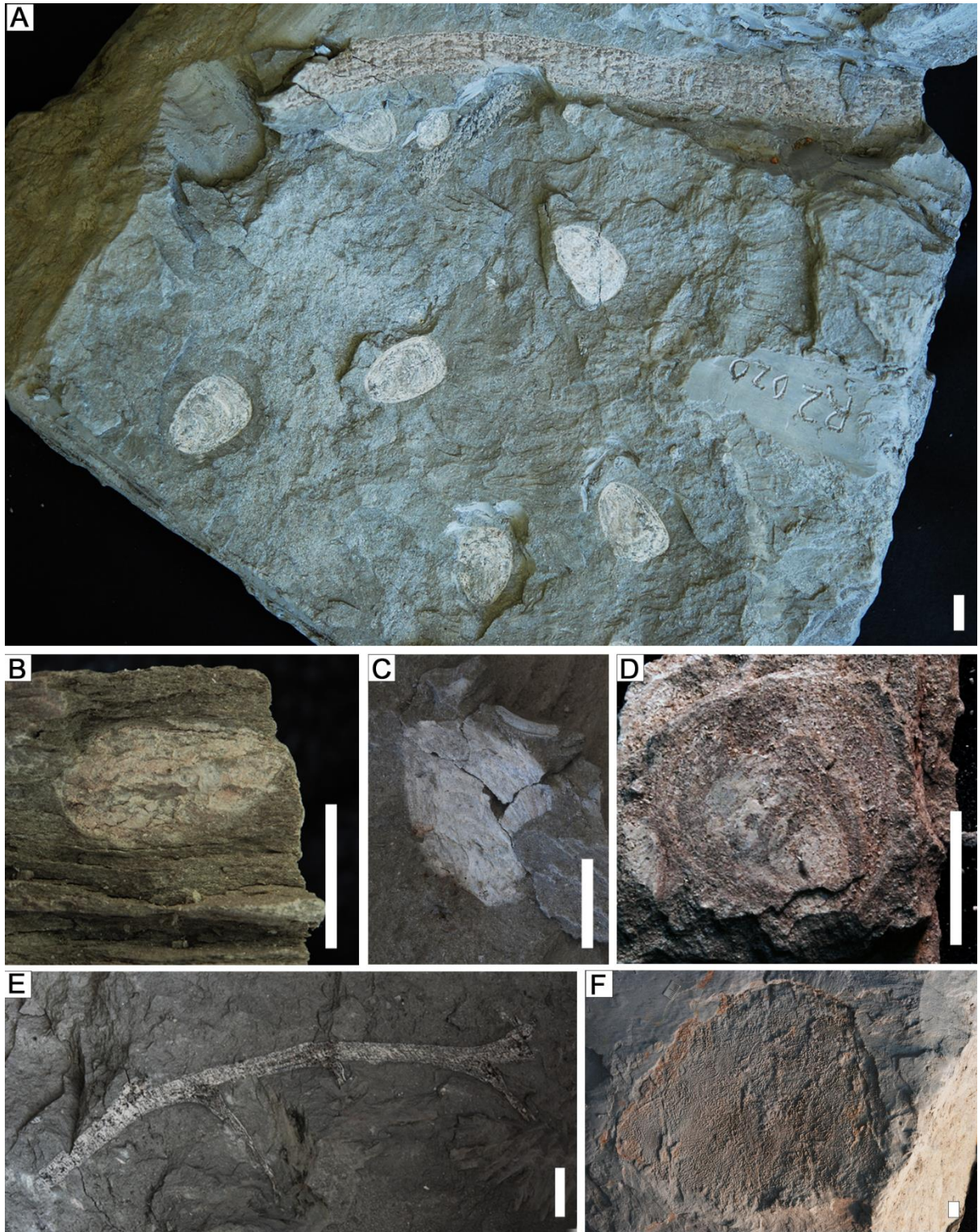


Figure 7.6: Body fossils from the Rabbit horizon. All scale bars = 1 cm A) A scatter of disarticulated bedding parallel lingulid brachiopod shells associated with a lycopod axis (AM 7824). B) A vertically-orientated, in situ lingulid brachiopod shell (AM 7838). C) An obliquely-orientated in situ lingulid brachiopod shell (AM 7839). D) Articulated lingulid valves in scissor position (AM 7836). E) An isolated 'woody' plant axis (AM 7837). F) Skull roof of an arthrodire placoderm, positioned anterior towards the top (AM 7835a).

7.5.2. Rabbit Ridge Depositional Model

Fluctuations in relative sea level are evident in the Rabbit Ridge succession. The lower section is interpreted as a 50 m thick, paralic, fluvial channel deposit with episodic marine influence. This gives way to a 25 m thick sandstone succession interpreted as a sheet-braided fluvial deposit. The overlying mudstone-dominated upper section represents a back-barrier lagoon. Collectively the succession evidences a long episode of overall progradation from the base of the lower section to the top of the middle section. Transgressive marine flooding of the coast would account for the overlying back-barrier depositional setting of the upper section. The recorded base-level changes may reflect an early Famennian eustatic signal.

A tidal channel to shallow-marine setting for the lower Witpoort Formation in the Eastern Cape (Hiller and Taylor 1992) inadequately explains the recorded facies progressions on Rabbit Ridge. Although not definitively excluded, we find no clear evidence favouring either of these depositional environments in the lower to middle sections. Rather, fluvial settings are proposed for the bulk of the above recorded arenaceous facies. Back-barrier deposits, developed during brief transgressions of the coastal plain, account for the development of mud-rich horizons and occurrence of marine trace and invertebrate fossils. This depositional model is congruent with recent, similar, reinterpretation of the Ordovician Peninsula Formation (Table Mountain Group, Cape Supergroup (Turner *et al.* 2011)).

7.5.3. Occurrence of Body Fossils

Invertebrate remains– Invertebrate body remains exclusively comprise lingulid valves, which are abundant and similarly distributed in the upper portions of the Rabbit horizon at both R1 and R2. At R1, lingulids occur in the upper half of the 9 m thick mudstone layer, but are extremely abundant in the upper metre. Many hundreds of valves were observed, orientated both oblique and parallel to bedding. The majority of shells are disarticulated whole valves, scattered on bedding planes with no clearly preferred orientation (Fig. 7.6A).

Fragments of broken valves, though common, are less abundant than entire valves. Four specimens were collected preserved in ‘scissor position’ with the valves articulated and slightly splayed (Fig. 7.6D). Among 123 whole collected valves, 34% are pedicle valves, 40% are brachial valves and the remainder are equivocal (Appendix 4B). Numerous valves at both R1 and R2 were preserved at an angle to bedding, from near horizontal to vertical (Fig. 7.6B, C). Oblique and vertical valves were found to be anterior upwards when collected *in situ* at outcrop and are interpreted as remaining in life position (Craig 1952; Pemberton and Kobluk 1978; Zonneveld and Greene 2010). Ubiquitous foreshortening during compaction (vertically orientated shells were reduced in length/width by up to 50%) had rendered many barely recognisable. No lingulid brachiopod valves were recovered in the lower part of the unit, however, a few lingulids, including bedding-parallel and oblique valves preserved in fine sand were discovered in the float from R1, suggesting that lingulids may also have occurred within sandy interbeds in the lower part of the Rabbit horizon.

Plant remains– Isolated plant axes occur in trough cross-stratified sandstone in the lower section, and in the lower of two thin argillaceous horizons in the middle section. These are all unbranched lycopod stems resembling ‘*Haplostigma*’ and, where preserved, leaf bases are round and helically distributed about the stem (Fig. 7.3G). Isolated plant remains occur throughout the Rabbit horizon at R1 and R2. These include undiagnostic pieces of lycopod cortex and unbranched lycopod axes up to 16 cm in length (Fig 7.6A) (usually preserving neither leaves nor fertile material), as well as alternately branched, coarsely mineralised non-lycopod axes (Fig. 7.6E). Plant remains at this horizon show a high degree of damage and fragmentation by comparison with those preserved at Waterloo Farm and Coombs Hill.

Vertebrate remains– The upper part of the Rabbit horizon preserves few putative bones, but one specimen with a fine, ripple-like dermal ornament is identified as an arthrodire placoderm skull roof (Fig. 7.6F). Sutures are not distinct but its general shape is compatible

and a preserved sensory canal junction (situated posteriorly on the left side) corresponds to the juncture between the main lateral line canal, the post marginal sensory canal and the infraorbital sensory canal. Ornamentation of the skull roof comprises an irregular pattern of miniscule steep-sided ‘ripples’, not observed in arthrodire bones previously described from the Witpoort Formation.

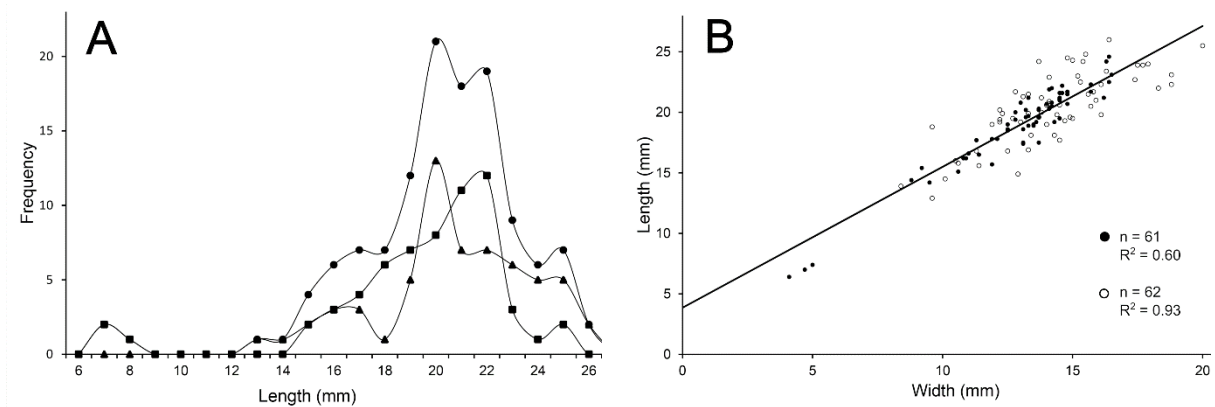


Figure 7.7: Size distribution of lingulid fossils. A) Size-frequency plot for lengths of lingulid brachiopod valves in the sample from the Rabbit horizon. Triangles represent data from R1, squares represent data from R2, circles represent the total sample. B) Scatter plot showing length versus width for the entire sample of lingulid brachiopod shells from the Rabbit horizon (n = 123). Open circles represent data from R1, full circles represent data from R2 and a line of regression for the whole sample is shown.

7.5.4. Size Distribution and Morphometric Variation of Lingulid Fossils

Size distribution within the lingulid sample was explored by graphically plotting valve lengths in size classes of 1 mm (Fig. 7.7A). The size–frequency curves for R1 and R2 are very similar and at both localities the most abundant valves are 20–22 mm in length. The total sample curve is bell-shaped and is strongly skewed to the right.

Variability in lingulid size and shape from R1 and R2 is assessed by graphically plotting length versus width in a sample of 123 whole, bedding-parallel specimens (Fig. 7.7B). A regression line for the entire sample with an r value of 0.89 ($r^2 = 0.78$) indicates a very strong correlation between these parameters. The level of consistency around the regression line

indicates a growth series belonging to a single species, here attributed to *Dignomia cf. lepta* (Appendix 4B).

The Rabbit Ridge lingulids are compared with stratigraphically well-constrained fossils from the R. D. F. Oosthuizen collection at Iziko Museum, Cape Town (Table 7-1). These include specimens from throughout the Early to Middle Devonian sequences represented by the Bokkeveld Group (Cape Supergroup) and overlying Wagen Drift Fm (lower Witteberg Group, Cape Supergroup), identified by Oosthuizen as '*Lingula*' *lepta*, '*L.*' *lamella*, '*L.*' *scalprum*, '*L.*' *keideli* and '*L.*' sp. (Oosthuizen 1984). Only small sample sets of complete valves were available for '*L.*' *lepta* (n=9), '*L.*' *keideli* (n=2) and '*L.*' sp. (n=8).

The mean lengths of those assigned to '*L.*' *lepta* and those of the Rabbit Ridge sample are very similar and their W:L and size ranges overlap considerably (Table 7-1). Specimens of '*L.*' *keideli* and '*L.*' sp. likewise have size and W:L ranges that fall within the range exhibited by the Rabbit Ridge sample, but are, on average, smaller (Table 7-1). There is, therefore, no clear proportional or size parameter permitting distinction of Rabbit Ridge *Dignomia* from other lingulid specimens reported from the Bokkeveld and Witteberg Groups, formerly identified as '*Lingula*' *lepta*, '*L.*' *keideli*' and '*L.*' indet.

Table 7-1: Comparison of Rabbit Ridge (RR) lingulids with Devonian, Cape Supergroup specimens from the Oosthuizen collection at Iziko Museum Cape Town.

Species	Number of specimens	Min length (mm)	Max length (mm)	Mean length (mm)	W:L	Mean W:L
<i>. lepta</i>	9	1.3	2.7	1.9	0.48–0.68	0.62
<i>L. keideli</i>	2	1.6	1.8	1.7	0.58–0.65	0.61
<i>L. sp.</i>	8	0.7	2.2	1.3	0.57–0.75	0.66
RR	123	0.6	2.6	2	0.57–0.84	0.69

7.6. Discussion

Lingulids exhibit remarkable conservatism in both morphology and lifestyle, which long resulted in the, now refuted (Emig, 2003), lumping of Devonian taxa within extant genera (Appendix 4B). Essential characteristics of Lingulaceae such as their ‘linguliform’ shape and chitinophosphatic shell have nonetheless remained unchanged since the Palaeozoic (Zabini *et al.* 2012). Studies of *in situ* fossil lingulid occurrences and their trace fossils moreover evidence concomitant continuity of burrowing and reburrowing behaviours (Craig 1952; Savazzi 1986; Zonneveld and Pemberton 2003; Zonneveld and Greene 2010). Studies of modern lingulids therefore provide a valuable analogue for informing analysis of ancient taxa, including not only structure, taphonomy and behaviour, but also ecology and geographic distribution (Emig 1997; Zonneveld *et al.* 2007).

Taphonomy– Extant lingulid valves have a low preservation potential being very brittle, and quick to decompose (Morse 1902; Kowalewski 1996; Emig 1997). For example, on tidal flats in Mexico, shells of *Glottidia* disarticulate soon after death, and are often completely eliminated within a few days, at most lasting a few weeks *post mortem* (Kowalewski 1996). As the shell composition of *Dignomia*, a thin shelled genus (Emig and Herrera 2006), was very similar to that of extant *Lingula* (Zabini *et al.* 2012) a similar taphonomic profile is expected. The commonly entire, unabraded condition of *Dignomia* shells at Rabbit Ridge, and *in situ* co-occurrences therefore indicate autochthonous, fairly rapid burial. The co-occurrence of whole valves with occasional broken fragments and a similar proportion of dorsal and ventral valves in the deposit furthermore suggest minimal sorting prior to burial. Apparent burial soon after death within the living environment permits meaningful extrapolation of population structure at Rabbit Ridge.

Population structure– Spatially discreet concentrations of lingulids at R1 and R2 (including both *in situ* and bedding parallel specimens) and their absence from several other

outcrops of the same horizon, indicates that the Rabbit Ridge community may have had a patchy distribution such as is commonly observed in modern lingulid populations (Kenchington and Hammond 1978; Kowalewski 1996; Zonneveld *et al.* 2007).

A linear regression curve (Fig. 7.7B) demonstrating strong correlation between length and width variables is consistent with a growth series of a single, monospecific population. We therefore assume a) that all specimens at Rabbit Ridge belong to a single species, and b) that the sample represents a natural death profile (with some time averaging).

The vast majority of shells (97%) are greater than half of the length of the largest in the sample. This indicates a death profile dominated by mature individuals as extant *Lingula* reach maturity at around 1/3 of their maximum size (Chuang 1961). Similarly, extant lingulid communities commonly exhibit populations of uniform-sized individuals with an absence of smaller representatives (Kenchington and Hammond 1978 and refs. therein). For example, 100 individuals of *Glottidia* dug from a mud-flat in California included no young (Morse 1902). Within living patches on tidal flats in Mexico, all individuals were found to be similar in size – representing a single cohort, which had simultaneously colonised the substrate (Kowalewski 1996). Likewise, 98 *in situ* Carboniferous fossil specimens collected from the Top Hosie Shale of Scotland constituted a bell-shaped curve with a paucity of small individuals (Craig 1952). Within extant populations this phenomenon has been explained by periodic failure of recruitment due to cannibalisation of free-swimming larvae by established populations (Kenchington and Hammond 1978, following Thayer 1975).

A bell shaped size-frequency distribution curve of fossils is indicative of a population of organisms continually occupying a habitat with favourable conditions (Rigby 1958). When skewed to the right (cf. Fig. 7.7A), such a curve indicates low rates of juvenile mortality (Rigby 1958; Kenchington and Hammond 1978). This suggests that the sample from Rabbit

Ridge represents a healthy population inhabiting a favourable environment, albeit with possible inhibition of subsequent cohorts.

Shells display a broad range of width to length ratios, but no greater than that observed in modern *Lingula* populations (Kenchington and Hammond 1978). The scientific literature provides little data on variability of this character in fossil populations, one exception being Ferguson (1963). Though width to length ratios are considered a useful taxonomic character in *Dignomia* species, only small samples sets are generally available (Emig and Herrera 2006). The unusually large sample from Rabbit Ridge provides a useful guide to understanding variability in a single *Dignomia* species.

Environmental controls– Lingulid brachiopods are known to thrive where competition is reduced due to environmental pressures (Zonneveld *et al.* 2007). The remarkable environmental tolerances of lingulids derive, in part, from their ability to rapidly withdraw down their burrows through contraction of an elongate fleshy pedicle anchored at the base (Morse 1902; Emig 1997). Positioned as such, a lingulid may remain dormant for several days, in a state likened to aestivation (Emig 1997; Zonneveld *et al.* 2007). Unusually amongst marine invertebrates, they can tolerate brackish water (though not freshwater habitats) (Hammen and Lum 1977; Hammond 1983) as well as extreme variability in a wide range of environmental parameters, such as temperature, pH, oxygen availability, bathymetric range (Emig 1997) and salinity (Morse 1902; Hammen and Lum 1977; Hammond 1983; Emig 1997). It is likely that lingulid dominance at the Rabbit horizon is related to competitive release in a shallow water habitat characterised by inhibitory environmental stresses. Environmental stresses in the Rabbit Ridge ecosystem may have included fluctuations in water temperature resulting from evident shallow water conditions at high-palaeolatitude, as well as stresses relating to oxygen concentration and salinity levels.

Although extant lingulid brachiopods are commonly reported from intertidal ‘mud flats’ (e.g. Kowalewski 1996; Darmarini *et al.* 2017), there has been some debate regarding their ability to inhabit argillaceous substrates. Grain size of host sediments appear to exert a strong control on lingulid distribution. Studies of modern (Paine 1970) and Mesozoic lingulids (Zonneveld *et al.* 2007) suggest that backshore muds may be unsuitable for lingulid burrow construction. Paine (1970) hypothesised that only below wave-base muds are habitable for lingulids, but tidal flat sediments comprising a dominant proportion of sub-sand sized clasts are unsuitable habitats. On this basis, he disputed an intertidal mudflat interpretation for *in situ* lingulid and bivalve associations in the Lower Carboniferous of Scotland (Ferguson 1963). Here, presenting an example of Famennian lingulids that have colonised shallow water backshore muds, we contend that Paine’s hypothesis is not necessarily predictive for Palaeozoic lingulid occurrences.

Present day lingulid occurrences are limited to equatorial and subtropical latitudes, within 40° of the equator (Emig 1997). A similar distribution has been suggested dating back to the Palaeozoic, and low temperatures are considered a limiting factor (Emig 1997; Zonneveld *et al.* 2007). However, occurrences in the Cape Supergroup suggest that lingulids were successful in a high-latitude setting throughout much of the Devonian Period. Living well south of 60°S during the Famennian Stage, the Rabbit Ridge lingulids would have experienced strongly seasonal, cool conditions. Amelioration of the Siluro-Devonian hothouse was underway precipitating, by the latest Famennian, the first confirmed continental glaciations since the end of the Ordovician Period (Isaacson *et al.* 2008). Limited solar radiation during winter, perhaps with sunless midwinters (Gess and Ahlberg 2018), may have caused low-temperature stress, particularly in the context of Famennian global cooling. Nonetheless, individual lingulids are inferred to have lived for up to a decade based on growth increments (Appendix 4B), and this is comparable to the lifespan of modern forms.

Gondwanan winter temperatures during this episode were therefore within their tolerances, yet perhaps inhibited potential competitors.

Evidence for low oxygen levels in the substrate is a common condition within Witpoort Formation mudstone horizons, which are generally black due to their high organic carbon content and at Waterloo Farm exhibit exceptional preservation of soft tissue (Gess and Whitfield 2020). Occurrence of ‘*Spirophyton*’ (helicoidal *Zoophycos*) at Rabbit Ridge and Waterloo Farm is furthermore consistent with low oxygen levels in the substrata (MacEachern *et al.* 2007). Such anoxic sediments are common in estuarine systems, even beneath oxygenated waters, as a result of high levels of organic input and low levels of oxygen diffusion into the fine-grained sediment (Day 1981a, b). Recent high-resolution geochemical study indicates cyclical alternations of anoxic and euxinic redox states within a mudstone horizon at Coombs Hill, perhaps related to seasonal fluctuations in freshwater input interacting with stratified bottom waters (Hoosen 2019). Relative tolerance of anoxia alone does not therefore account for the presence of lingulids at Rabbit Ridge in contrast to infaunal bivalves at Coombs Hill (see below).

Based on the depositional setting, seasonal salinity fluctuations in the water column are likely to have occurred. Extant *Lingula* are able to tolerate dramatic reductions in salinity (half or less compared with normal seawater concentration) during short intervals such as tidal fluctuations or for episodes up to several days, but longer periods of exposure to reduced salinities may cause death or impairment (Hammen and Lum 1977; Hammond 1983; Emig 1997). Reporting on a monospecific lingulid accumulation from the lower Witteberg Group, which stratigraphically underlies the Witpoort Formation, Hiller (1990, p. 225), therefore hypothesised that “abundant *Lingula* in the Witteberg Group represents a nearshore setting, possibly with fluctuating salinity”. Observations of modern and fossil lingulids indicate that extreme salinity reduction is a common cause of lingulid die offs (Morse 1902; Craig 1952;

Kowalewski 1996; Emig and Gutiérrez-Marco 1997; Aceñolaza *et al.* 2003; Emig and Herrera 2006). Horizons with abundant bedding parallel valves, such as those associated with plant axes at Rabbit Ridge (Fig. 7.6A) may consequently have resulted from episodes of increased freshwater contamination.

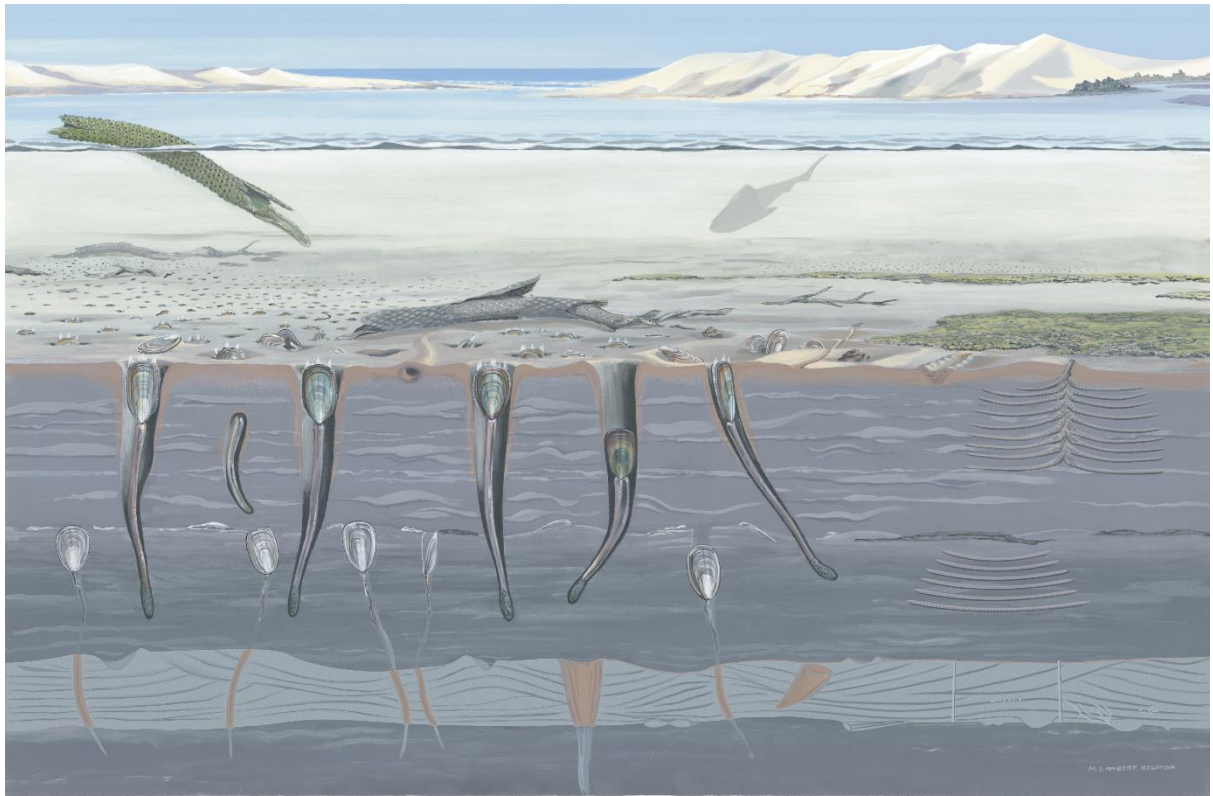


Figure 7.8: Palaeoenvironmental reconstruction of the Rabbit Ridge lingulid bed, modified from painting by Maggie Newman.

Globally, the Late Devonian was a time of stressful marine conditions punctuated by a series of ‘extinction events’ associated with atmospheric and oceanic perturbations and rapid sea-level fluctuations (Bond and Wignall 2008; Carmichael *et al.* 2019; Percival *et al.* 2019). The early Famennian age of the reported lingulids potentially places them in the aftermath of the Kellwasser Event (ca. 372 Ma). Although no marker bed of this event has been recognised in the Witteberg Group it is likely, following Cooper’s (1986) event stratigraphy model, to approximately correspond with the Weltevrede–Witpoort Formation contact some

one hundred metres underlying the Rabbit Horizon. This suggests that the Rabbit Ridge fauna significantly postdates the Kellwasser Event, making it unlikely that the assemblage represents a disaster fauna. Moreover, the population does not exhibit any Lilliput effect (size decrease of organisms) relative to congeneric examples from underlying strata – *contra* dwarfism observed in lingulids that proliferated globally following the Permo-Triassic mass extinction (Rodland and Bottjer 2001; Peng *et al.* 2006).

Depositional setting of the lingulid bed– The lingulid-bearing mudstone horizon on Rabbit Ridge is interpreted as a variant of the back-barrier depositional systems described at other argillaceous horizons in the Witpoort Formation (Hiller 1990; Hiller and Taylor 1992; Harris *et al.* 2021). The extensive mudstone lens, bearing a marine biotic signature and extending parallel to the inferred palaeo-shoreline indicates a sheltered, wave-dominated lagoonal setting (Fig. 7.8). This likely derived from transgressive submergence of a low-gradient coastal plain (see above). The lagoonal substrate hosted a monospecific lingulid population associated with a suite of marine trace makers. Neither the lingulid fossils (Table 7-1), nor associated ichnofossils bear evidence of the diminution normally characteristic of brackish water faunas (Bann *et al.* 2004; Buatois *et al.* 2005).

During the early stages of backshore flooding and barrier development, high energy depositional events periodically interrupted mud settling. The resultant wave-rippled and parallel-laminated interbeds were colonised by infaunal and epifaunal marine trace makers. Resultant traces are dominated by *Nereites*, an opportunistic infaunal grazing trace common in Palaeozoic shallow water and intertidal settings (Mángano *et al.* 2002b). The occurrence of ‘*Spirophyton*’ in the muddy back-barrier environments of Rabbit Ridge and Waterloo Farm has parallels with occurrences of *Spirophyton* and *Zoophycos* in Late Devonian strata from eastern New York State. These latter are interpreted as being representative of opportunistic colonisation of ephemeral pools on estuarine floodplains (Miller 1991). During the

Palaeozoic, *Spirophyton* and *Zoophycos* trace makers commonly occupied shallow-marine deposits, though they subsequently became restricted to deeper marine environments (Bottjer *et al.* 1988).

By contrast with Rabbit Ridge, fossil localities at Waterloo Farm and Coombs Hill, which are stratigraphically higher in the Witpoort Formation, include no invertebrate species with clear marine signatures. At Waterloo Farm, estuarine black shale facies of the upper Witpoort Formation (Fig. 7.1) preserve a rich and diverse accumulation of vertebrate bones and soft tissues, eurypterid arthropods, bivalve molluscs, plants and algae, collectively indicative of a brackish water setting connected to both marine and fluvial habitats (Gess and Whitfield 2020). Its utilisation by marine, non-marine and resident fish taxa, in particular as a nursery environment, reflects strategies evidenced in the fossil record and still prevalent today (Gess and Whitfield 2020). At Waterloo Farm hundreds of preserved bivalves represent a single species of *Naiadites*, a characteristically fresh to brackish water epifaunal genus (Huwe 2006, Scholze and Gess 2017). Coombs Hill, located about a kilometre east of the R0 Rabbit Ridge outcrop exposes an 80 m thick quartzitic package containing a series of lagoonal mudstone deposits, from higher within the Witpoort Formation sequence. The mudstones are rich in bivalves, which are dominated by autochthonous remains of a probable infaunal taxon, and associated with patchily distributed, unusually intact plant remains (Harris *et al.* 2021). Like Waterloo Farm, Coombs Hill appears to represent a more proximal, less marine facies than Rabbit Ridge.

The Late Devonian (Frasnian to Famennian) Catskill Formation of New York State similarly exhibits an increasingly continental signal up-succession, with lingulids in the lower part of the formation giving way to freshwater unionid bivalve molluscs (*Archanodon catskillensis*) (Broussard *et al.* 2020). This is consistent with a progradational model for deposition of the Catskill Formation (Broussard *et al.* 2020). It is, likewise, plausible that the

Witpoort Formation records an overall progradational system in which lingulids, reflecting more marine facies are replaced up-sequence by bivalves in more freshwater influenced lagoonal or estuarine settings. Indeed, the lingulid brachiopods at Rabbit Ridge are the stratigraphically highest unequivocally marine invertebrates known within the Cape Supergroup.

7.7. Conclusions

1. Lingulid brachiopods are, for the first time, reported from a horizon within the Witpoort Formation (Witteberg Group, Cape Supergroup). This provides the youngest record of a characteristically marine invertebrate within the Cape Supergroup. Depauperate shallow-marine faunas containing lingulids, from the underlying Weltevrede Formation have previously provided the upper limit of marine invertebrate taxa.
2. They occur superabundantly, preserved within their living environment in a backshore setting postulated as a back-barrier lagoon.
3. Lingulids were patchily distributed within their habitat and their population structure indicates a favourable environment with low rates of juvenile mortality.
4. The monospecific nature of the invertebrate body fossil assemblage results from lingulids unparalleled ability to exploit stressful marginal environments.
5. The new material provides the youngest known record of the widespread genus *Dignomia* which, despite occurrence in the Frasnian-aged upper Weltevrede Formation of South Africa, has generally been considered to have only survived from the Ordovician to the mid-Devonian.

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CHAPTER 8 – DISCUSSION AND CONCLUSIONS

The results of this thesis give, for the first time, a detailed account of sedimentary facies throughout the stratigraphic column of the Witpoort Formation in the Eastern Cape, and include evidence from sediment transport patterns, ichnology and body fossil occurrences. The provenance of each data point is recorded in stratigraphic columns in the Appendices to this work, and it is hoped that the model here proposed will be rigorously tested and refined with higher resolution analyses utilising a variety of methods. The main results are summarised below and their broader implications for Devonian studies subsequently explored.

8.1. Stratigraphic framework

The studied outcrop includes two composite sections of the Witpoort Formation in the Eastern Cape. A ~360 m thick reference section from Waaipoort (Steytlerville area) cross-sections the Witpoort Formation from its lower contact with the Weltevrede Formation to its upper contact with the Kweekvlei Formation. About 200 km east, in the ‘eastern sector’, several sections from within a 40 km radius of Makhandia are assembled as a composite section. According to the stratigraphic framework presented in Chapter 4, the most prominent shale-rich intervals in the Witpoort Formation succession are traceable for hundreds of kilometres, marking regionally correlatable stratigraphic surfaces. It is, furthermore, argued that the three members of the formation extend across the study area. In lieu of the secondary mineral staining that separates the Rooirand Member from its overlying Perdepoort Member further west, these members are here recognised in the eastern sector based on a set of sedimentological criteria. Westward from Waaipoort, the members have long been recognised by a sharp colour contrast between the 270 m thick reddish-brown quartzite of the Rooirand Member and the 60 m thick white quartzitic sandstone of the overlying Perdepoort Member (SACS, 1994; Johnson *et al.* 2006). The Skitterykloof Member is a thin (17–20 m

thick) unit at the top of the Formation that is defined by the presence of gravel clasts. The reddish tint of the Rooirand Member is derived from precipitation of remobilised metals along fractures, whereas the Perdepoort Member retains a whitish colour by the lack thereof (Olivier *et al.* 2011). East from Waaipoort, secondary staining is increasingly prevalent throughout the formation, which inhibits a clear separation of these members (Olivier *et al.* 2011).

Across the study area the Rooirand Member, comprising about 75% of the formation's thickness, overlies the Weltevrede Formation with a sharp transition from heterolithic mudrock-dominated to sandstone-dominated strata. The Rooirand Member is separable from the Perdepoort Member by its cyclical sedimentation patterns, which include erosively bounded muddier-upwards units with intermittent conglomerate and mudrock, the latter often with concentrations of fossils. The Perdepoort Member is distinctive with its extensive, tabular quartzites and the paucity of mudrock, bioturbation or intraclasts. The Skitterykloof Member represents the transition from the Perdepoort Member to the Kweekvlei Formation shale. It is a conglomeratic unit with interfingering arenaceous and argillaceous facies, and abundant bioturbation. In the eastern sector a 'Skitterykloof Member equivalent' is described from a section at Alicedale. It lacks the lithic pebbles that define the member in the Western Cape (SACS 1994), but contains intraformational conglomerate and interbedded, bioturbated mudrock, representing a similar 'transition zone' (Loock 1967) between the arenitic Perdepoort Member and the shaly Kweekvlei Formation as found at Waaipoort.

Variable thicknesses that have been proposed for the Witpoort Formation in this eastern sector are evidently the result of structural thickening. For example, E. D. Mountain's figure of 730 m, measured in the Coombs Valley (cited in Johnson 1976), gives no account of structural repetition of strata. Tight folding with a negligible change in dip angle was documented in a freshly exposed road cutting at the Coombs Hill fossil locality bordering the

Coombs Valley (Harris *et al.* 2021). In the valley itself, numerous changes in younging direction suggest that forming a complete stratigraphic section of this formation is unlikely there (pers. obs.). Similarly, D.K. Toerien's estimated thicknesses of 850 m near Alicedale (pers. comm. in Johnson 1976) is likely affected by cryptic duplications, as contended by Meyer (1965). The estimate of 350 m given by Rhodes University structural geologist R. W. Harris (cited in Hiller and Taylor 1992) appears to be more accurate. It is supported by similar scales of fluvio-estuarine cyclicity across the study area. This result suggests that the basin underwent uniform rates of subsidence, which is in accordance with conceptions that the Witteberg Group represents an interval of basinal stabilisation (Tankard *et al.* 2012). Apparent thickening of the stratigraphy to the east of the Cape Fold Belt is, therefore, a result of post-depositional, structural thickening rather than syn-sedimentary tectonic effects (Booth 2011).

8.2. Depositional environments

Based on vertical and lateral relationships, body and trace fossil content and flow vectors of the 12 sedimentary facies described in Chapter 5, two principal facies assemblages are recognised, the first being braided river deposits (FA 1) and the second being estuarine deposits (FA 2). The latter are subdivided in four sub-environments, which indicate a spectrum from inner to outer estuary sediments. Open-marine deposits were not recognised in the studied strata, in contrast to previous thinking on the Witpoort Formation, which posits a shoreface, beach and barrier island and lagoonal deposit (Thamm and Johnson 2006).

The facies assemblages, described in Chapter 5, alternate in repetitive cycles in the Witpoort Formation that are on the order of metres to tens of metres thick, which typically show a progression from the most landward facies at the base, passing upwards to the more marine influenced. This cyclicity is most evident in the Rooirand Member, wherein both of the principal facies assemblages (FA 1 and FA 2) occur in greater or lesser proportions within

each cycle, or cyclothem (*sensu* Catuneanu 2006). Sedimentary cyclicity is less clear in the Perdepoort Member, which is exclusively fluvial (FA 1). The Skitterykloof Member contains both FA 1 and FA 2, evidencing rapid, non-cyclical base-level fluctuations, the significance of which is discussed below in this section 8.6. Selected stratigraphic log examples from the study area are examined here to show the cyclical variations (Fig. 8.1 A–G).

A typical Rooirand Member cycle is initiated by a prominent lag deposit, often with a medium–thick bed of intraclast conglomerate (Facies Gi) containing reworked, up to boulder-sized clasts from the top of the previous cycle. This lag deposit may be localised and is not evident at the base of every cycle. The facies interpretation given in Chapter 5.2.1 asserted that the clasts would have originated within the immediate depositional environment without significant transport distances. They often contain sub-rounded, deformed sand clasts, for which no better explanation than cryogenic consolidation and transport has been found (cf. Reinson and Rosen 1982; Runkel *et al.* 2010). Alternative explanations of early chemical cementation or algal binding (Pflüger and Gresse 1996) were found to be unsatisfactory. This lag deposit represents a sequence boundary (in the sense of a genetic sequence, *sensu* Catuneanu *et al.* 2009), marking a sedimentation hiatus and forestepping of the depositional environment. The lag deposit is, in many cases, overlain by sheet-braided fluvial deposits, which are highly amalgamated, cross-stratified and structureless sandstone (FA 1) forming accumulations of between 1.5 and 8 m thick in the illustrated examples (Fig. 8.1A–F). The fluvial deposits are, in turn, retrogradationally overlain by estuarine deposits (FA 2a, 2b, 2c, normally in that order). The estuarine facies assemblage differs from the fluvial one by evidence for tidally-influenced channel processes, specifically the deposition of fine-grained sediment, with mud-draped sandstone laminae (and sometimes double drapes) and lenticular bedforms with ripple cross-laminated, silty toe sets. An increase in mud content upwards produces a heterolithic transition zone (FA 2b) and the accumulation of thick mudrock

deposits (FA 2c). Brackish water trace fossil assemblages in FA's 2b and 2c indicate elevated salinities in the deposits (see Chapter 6.3). In many cases, thick accumulations of brackish water mudrock (FA 2c) contain the maximum flooding surface preserved in a cycle, and are overlain by the next sequence boundary and a return to FA 1 deposition (Fig. 8.1C, D and E). In some successions, however, muddy basin deposits have been removed, and are only evidenced by reworked clasts above the sequence boundary (Fig. 8.1B). Certain stratigraphic horizons from the lower Rooirand Member at Waaipoort preserve dune deposits that resemble the fluvio-estuarine channel facies assemblage (FA 2a), but show a landwards accretion direction. They are interpreted as outer estuary deposits (FA 2d). Flow vectors are the only clear criterion presently separating them from inner estuary, bayhead delta sandbars, which have a strongly offshore flow signal. FA 2d was not recognised in the eastern sector of the study area.

The described erosionally-bound, retrogradational cycles of the Rooirand Member are 5–12 m thick in six of the nine illustrated examples (Fig. 8.1A–E). They reach up to 21 m thick in the study area (i.e. WP 1, m. 89–110). Some successions, however, exclusively comprise estuarine deposits that are stacked in complex successions tens of metres thick, as in the lowermost and uppermost strata in the Rooirand Member (Fig. 8.1A and F). On the other hand, accumulations of FA 1 reaching 30–50 m in thickness are prevalent in the middle part of the Rooirand Member, where intraclast stringers at intervals of 10–25 m provide evidence of reworked mud deposits. The Perdepoort Member contains up to 80 m of supermature FA 1 sheet sandstone that is developed in laterally extensive tabular units of 10–20 m thick and hundreds of metres in lateral extent, separated by slight accumulations of fines that form prominent notches in the outcrop (Fig. 8.1G). By their thickness these units would appear to represent a similar order of cyclicity to that recognised in the Rooirand Member, but with only the fluvial component preserved.

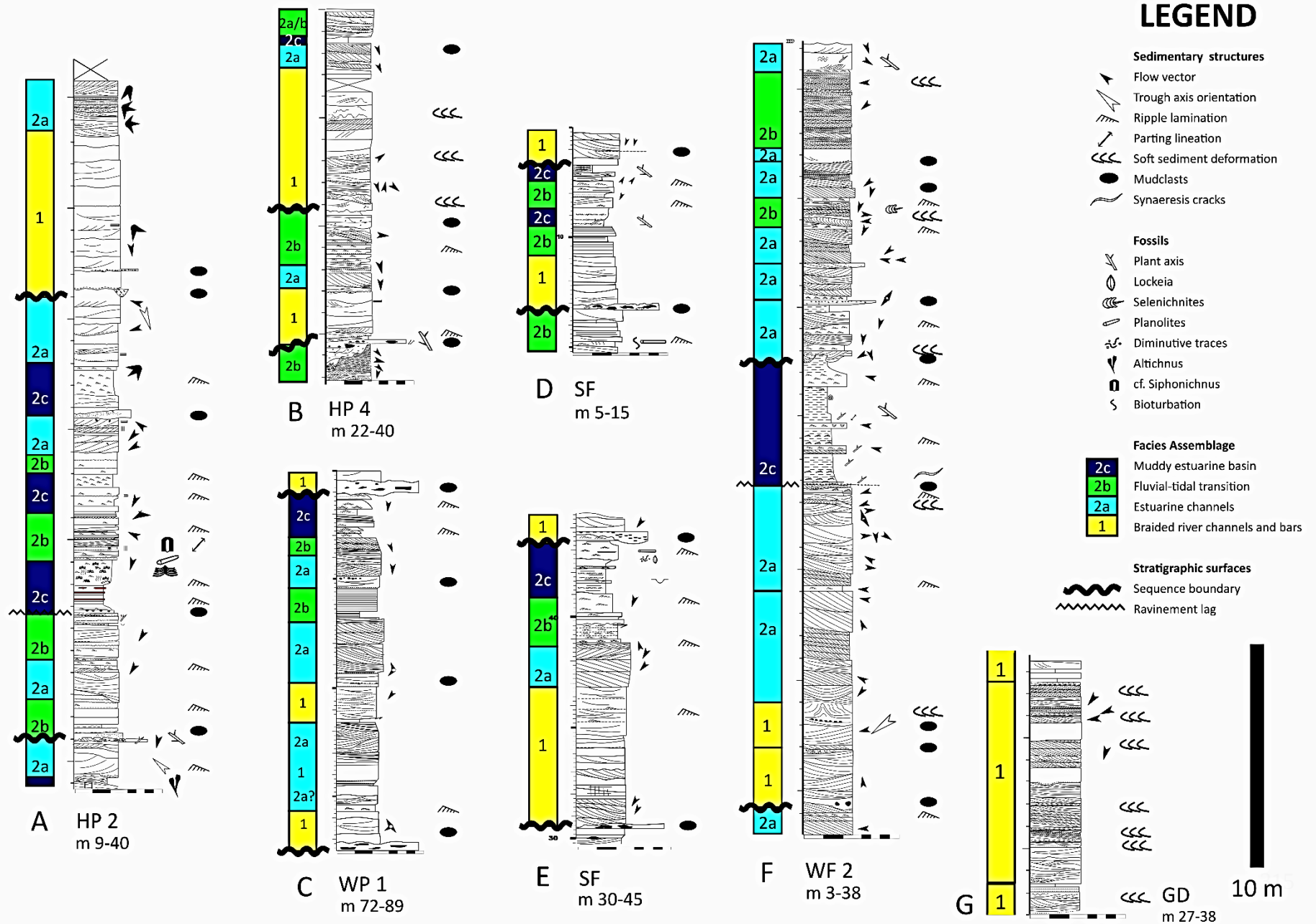


Figure 8.1: Stratigraphic columns showing sedimentary cyclicity in the Witpoort Formation, A–F) Rooirand Member. G) Perdepoort Member. Section labels are as given in previous chapter.

The metre-scale cyclicity in the Rooirand Member is primary evidence of base-level change. It shows a repeated pattern of erosional incision followed by lowstand fluvial deposition, passing upwards to fluvio-estuarine channels, bayhead deltas and culminating in muddy estuarine basin deposits or, in places, outer estuary sandbars. This retrogradational succession is typical for estuarine facies models (Dalrymple *et al.* 1992; Dalrymple and Zaitlin 1994; Plint and Wadsworth 2003; Fielding *et al.* 2020; Mángano *et al.* 2023).

Estuaries have a tripartite distribution of facies, which includes high-energy facies landward (inner estuary) and seaward (outer estuary) of a relatively low-energy central zone, alternatively named the estuary funnel (Nichols *et al.* 1991) or the estuary central basin (Dalrymple and Choi 2007). This is due to bedload convergence that occurs in the middle portion of estuaries (Dalrymple and Choi 2007). Wave-dominated and tide-dominated estuaries can be distinguished from their characteristic deposits (Dalrymple *et al.* 1992). Wave-dominated estuaries produce a distinctly mud-rich central basin (Zaitlin and Shultz 1990; Nichols *et al.* 1991), whereas tide-dominated estuaries tend to be more heterolithic (Allen and Posamentier 1993; Dalrymple and Choi 2007). The very bimodal separation between sand- and mud-dominated facies in the Witpoort Formation points to wave-dominated estuaries, since low-energy estuarine central basin muds indicate deposition in the former class (Roy *et al.* 1980; Zaitlin and Shultz 1990; Nichols *et al.* 1991; Mackay and Dalrymple 2011; Gandini *et al.* 2017). Accumulation of homogenous muds, for example, may be explained by periodic closure of the mouth, forming a barred estuary (Roy *et al.* 1980). The spectrum of wave and tide influence in estuaries is the result of interplays between these factors. Barred estuaries generally reflect coasts with smaller tidal ranges (Nichols *et al.* 1991; Dalrymple and Zaitlin 1994; Roy *et al.* 2001; Perissinotto *et al.* 2010), but do not necessarily prove a microtidal regime (Davis 1994; Vakarelov *et al.* 2012).

Incompleteness of tidal rhythmites in the study area would be accounted for by a microtidal, wave-dominated regime, which is less likely to form well-developed neap-spring-neap rhythmicity than mixed wave and tide or tide-dominated systems (Choi and Park 2000; Dalrymple and Choi 2007). Though microtidal regimes are sometimes expected for high-latitude environments due to a decrease in the tidal bulge towards the poles, this is not always the case. In fact, tidal refraction and amplification can cause macrotidal ranges in polar environments such as the modern day Weddel Sea (Dalrymple and Padman 2018). A concerted investigation into tidalites in the Witteberg Group may test whether a microtidal regime existed in the Agulhas Sea during the Late Devonian, as suggested by the present results.

Estuaries, by definition, form on transgressive coasts, reflecting marine flooding of an abandoned delta plain or a fluvially-incised valley (Dalrymple 2006). Erosionally-bound retrogradational cycles such as those described for the Rooirand Member are consistent with incised-valley fill (IVF) successions (Plint and Wadsworth 2003; Wang *et al.* 2019). They are distinguished from simple channel fills by the stacking patterns from fluvial sheetwash, upwards to tidally-modulated channels and estuarine muds. This retrogradation can produce an increasing scale of bedform upwards in a cycle, as seen in the study area. Simple channel fills, by contrast, typically have a thinning-upwards facies progression, reflecting shallowing of the channel (Gibling 2006; Miall 2010). Incised-valley fills average between 20 m and 50 m thick in the rock record, depending largely on the rates of subsidence and sea-level rise, and extend for kilometres to tens of kilometres wide and up to hundreds of kilometres long in a shore-perpendicular direction (Wang *et al.* 2019). The present analysis does not permit a clear understanding of the scale of the inferred IVFs. The described cyclicity is thinner, on average, than is expected for IVFs, and may represent small base-level changes within larger valley fill sequences (Zaitlin *et al.* 1994). The delimitation of incised-valley fills in the rock

record may be achieved by a combination of facies analysis and subsurface data (geophysical and/or core), with the identification of regionally significant erosional surfaces that mark abrupt basinward shifts in depositional facies (Plint and Wadsworth 2003).

8.3. Shoreline orientation

Flow vectors from the study area were analysed according to facies assemblage, and the resulting interpretation is that the modal flow patterns for all Witpoort Formation stratigraphic sections in the study area approximate basinwards transport. Plotting these modal orientations on a map of the study area gives an indication of the shoreline orientation. The Rooirand Member in the eastern sector of the study area exhibits a SE flow trend, whereas in the west of the study area the trend is SSE (Fig. 8.2A). The Perdepoort Member shows a WSW modal flow trend across the study area (Fig. 8.2B). The results are compared with a diagram from Theron and Looock (1988) showing the generalised flow vectors for the Witteberg Group (Fig. 8.2C). The trends illustrated by Theron and Looock (1988) are strongly comparable with the Rooirand Member data in the present analysis. The more westerly trend of the Perdepoort Member may explain the westward trends reported by Johnson (1991), which he interpreted as representing alongshore sediment transport in open-marine conditions. For the present analysis, the slight westering of the modal flow trend of the Perdepoort Member may indicate a change in progradation direction and shoreline orientation during its deposition.

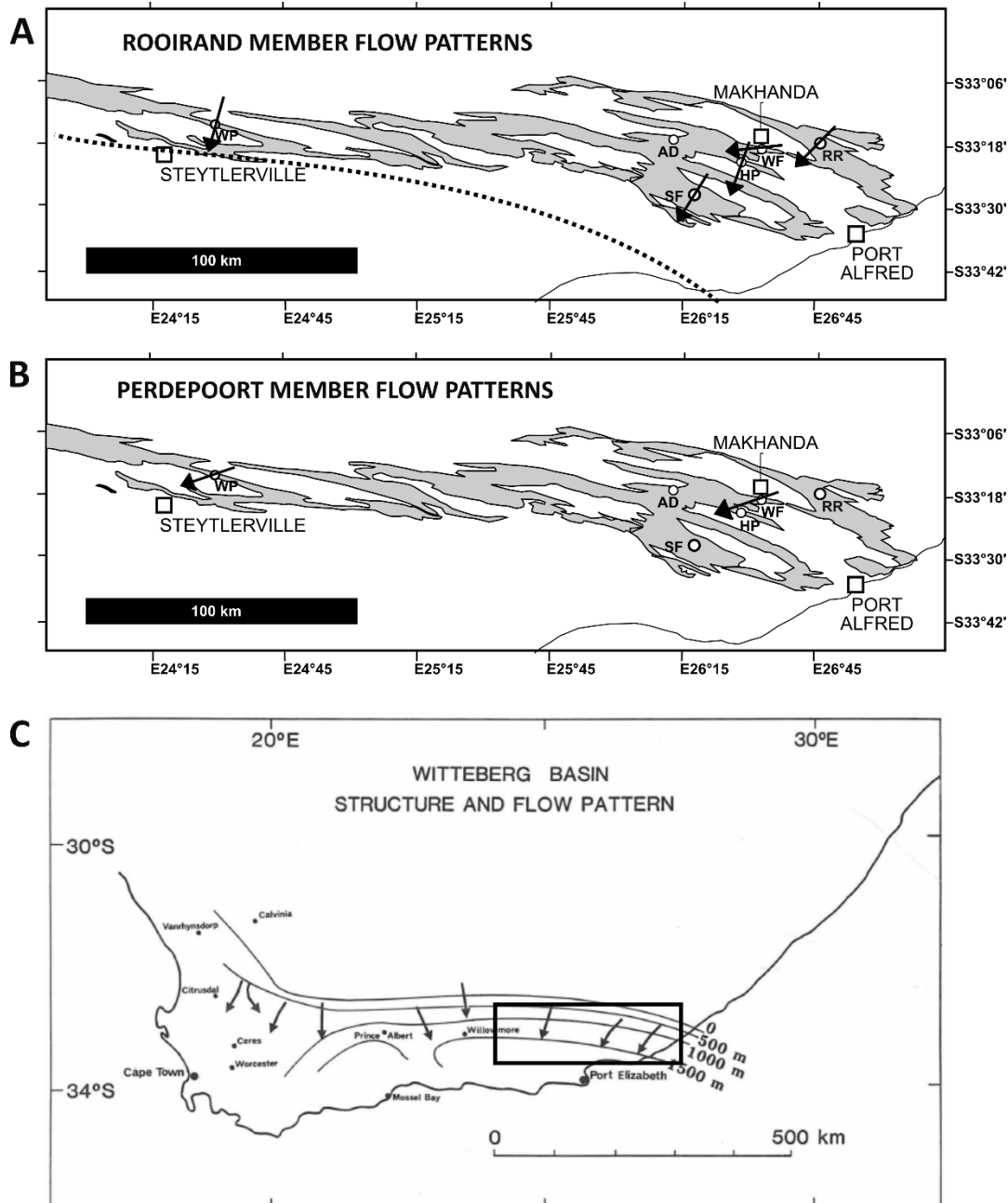


Figure 8.2: Flow patterns (black arrows) of the Witpoort Formation in the study area. A) The Rooirand Member, with a stippled line indicating the estimated shoreline orientation during maximum transgression. B) The Perdepoort Member, where the shoreline is interpreted as having occurred westward of the study area. C) Flow trends for the Witteberg Group from Theron and Looek (1988, their Fig. 9), with the approximate location of the study area outlined.

8.4. Sequence stratigraphy

This new depositional model for the Witpoort Formation will have fundamental implications for a sequence stratigraphic interpretation, which has not been attempted previously at sub-formation scale. Were an open-marine genesis for the mature arenitic strata

that dominate the stratigraphic column accepted, along with a backshore lagoonal or estuarine provenance for the mudrock interbeds (that is well accepted based on palaeontological grounds), shoreface or beach deposits (cf. Thamm and Johnson 2006) overlain by lagoonal muds would invoke shoreline progradation (regression), whereas burial of a backshore mudrock by a marine sandbody would imply backstepping (transgression) of the coast. The present analysis has, however, inverted the previously accepted model. Sandstone-dominated strata are shown to have originated on a coastal braid-plain and the development of estuary-basin mudrock therefore indicates transgressive flooding of the coast. Moreover, the sequence stratigraphy of non-marine or paralic systems differs from that of open-marine shelfal systems, particularly in the correlatability of stratigraphic cyclicity. The former is more likely to have been influenced by depositional hiatuses due to fluvial erosion, or subaerial unconformities (Catuneanu 2006), and cyclicity may therefore reflect local variation to a greater extent than in a corresponding shelfal system.

Controls on base-level change in the Witpoort Formation may be explained with reference to autogenic processes like subsidence rate and sediment supply, or allogenic controls, which include climate and eustasy (Blum and Törnqvist 2000; Saha and Bhattacharya 2023). Many non-marine sequence stratigraphic studies emphasise either allogenic or autogenic forcing, although both mechanisms likely played interrelated roles. For example, the rate of sediment supply is affected by climate change (Saha and Bhattacharya 2023), which in turn controls eustasy to a great extent (Blum and Törnqvist 2000). Subsidence rate too may be affected by allogenic processes, such as loading by water mass during a transgression (Plint and Wadsworth 2003), or isostatic adjustment following glaciation and deglaciation (Belzyt *et al.* 2021). Interconnected, and often counterintuitive, feedbacks between allogenic and autogenic controls on base-level are emphasised by Blum and Törnqvist (2000), with a theoretical example of the effects of glacial loading on accommodation space, where increased isostatic

subsidence would be countered by falling sea levels. A recent example of this negative feedback loop is the relative sea-level rise recorded in Europe during the Younger Dryas glacio-eustatic regression about 12 000 years ago, which is explained by increased subsidence due to glacial loading (Fjeldskaar and Amantov 2018).

Cyclical facies changes at a scale of metres to tens of metres (5–21 m thick) in the Rooirand Member are interpreted as reflecting short term base-level change. The cycles are underlain by erosional surfaces that indicate the minimum of accommodation. This surface likely encompasses the falling stage of a cycle in base-level change, which is represented by a hiatus in deposition (Catuneanu 2006). In the sense of a ‘depositional sequence’, the basal erosion surface records a sequence boundary. Overlying fluvial strata are interpreted as representing lowstand deposits, as is typical for the lower portions of many incised-valley fill successions (Plint 1983; Allen and Posamentier 1993; Zaitlin *et al.* 1994; Blum *et al.* 2013; Mángano *et al.* 2023). Transgression produces a muddier-upward trend, with the maximum flooding surface (the most landward position of the shoreline; Zaitlin *et al.* 1994) usually recorded in the development of mudrock facies in the study area, which have been deposited within the zone of maximum turbidity in the estuary basin. This situation resembles the middle section of a typical IVF that, depending on a multiplicity of variables, can occur at a significant distance from the coastline during maximum transgression (cf. segment 2 of Zaitlin *et al.* 1994). These interpretations for the recorded cyclic facies changes in the Witpoort Formation present a model for testing the controls on base level.

The relative significance of the various stratigraphic surfaces recognised in the succession can only be superficially evaluated with the present dataset. As previously mentioned, recognition of the scale of incised valleys in the rock record relies on a good understanding of dip and strike variations in sedimentary facies, and identification of regionally significant surfaces or erosion (Plint and Wadsworth 2003; Miall 2016).

8.5. Famennian event stratigraphy

At the stage level, the Famennian is widely regarded to be a time of global sea-level fall (Johnson *et al.* 1985, 1986; Becker *et al.* 2012; Haq and Schutter 2008; Marcilly *et al.* 2022; Van der Meer *et al.* 2022). This second-order eustatic regression corresponds to an interval of global cooling beginning at around the Frasnian–Famennian (F–F) boundary, inferred from isotopic records (Joachimski *et al.* 2009), which culminated in widespread continental glaciation in the latest Famennian (Isaacson *et al.* 2008; Brezinski *et al.* 2009, 2010; Lakin *et al.* 2016, 2021). Evidence for Devonian glaciation predating this time is tenuous (Lakin *et al.* 2016) and the question of Devonian glacioeustasy is an open one (Brett *et al.* 2011; McClung *et al.* 2013; De Vleeschouwer *et al.* 2014).

Interpretation of higher frequency sea-level change in the Famennian is complicated as it is affected by interpretations of facies change, variable rates of subsidence and the influence of syn-sedimentary tectonics (that affects the stratigraphy of many successions of this time interval). Evidence for syn-sedimentary tectonic activity affecting Late Devonian stratigraphic successions has been described for the Moroccan Anti-Atlas (Wendt 2021), the Canning Basin of Australia (Chow *et al.* 2004), the South China Basin (Zong *et al.* 2016) and the Appalachian Basin (Smith and Jacobi 2001). Nonetheless, eustatic signals are reflected in all these successions, possibly because sea-level fluctuations are more rapid than tectonic events.

Correlation between many widespread sections is achieved with event stratigraphy and biostratigraphy. Anoxic black shale events occurred throughout the Late Devonian. Important ones that are recognised in many basins include the Kellwasser Event, which marks the F–F boundary (Bond and Wignall 2008; Carmichael *et al.* 2019), the mid-Famennian *annulata* Event (Racka *et al.* 2010; Hartenfels and Becker 2016), the late Famennian Dasberg Event (Hartenfels and Becker 2009; Myrow *et al.* 2011) and the terminal Famennian Hangenberg

Event (Caplan and Bustin 1999; Myrow *et al.* 2014; Kaiser *et al.* 2016). These black shale horizons are widely thought to represent eustatic transgressions (Caplan and Bustin 1999; Bond and Wignall 2008; Racka *et al.* 2010; De Vleeschouwer *et al.* 2013; Myrow *et al.* 2014; Becker *et al.* 2012; 2020; Kumpan *et al.* 2015; Hartenfels and Becker 2016; Kaiser *et al.* 2016; Zhang *et al.* 2019; Ma *et al.* 2020; Wendt 2021; Hartenfels *et al.* 2022). Some studies, however, have proposed regression and restriction of epicontinental seas for the development of marine anoxia during some of these events (Bábek *et al.* 2016; Dopieralska *et al.* 2016). Thus, there is no clear consensus about sea-level changes, even in the most intensively studied intervals. Based on the evaluations of most research, these events are interpreted below as transgressive.

The Late Frasnian Stage is thought to have produced significant transgressions with perhaps the highest sea levels of the Devonian Period (Bond *et al.* 2013; Becker *et al.* 2020; Wendt 2021). Late Frasnian transgression culminated in the biphasic anoxic episode of the Kellwasser Event, producing a pair of anoxic black shales recognised in many marine shelf successions (Carmichael *et al.* 2019) and also, less commonly, in shallow-marine deposits (Bond *et al.* 2013; Mottequin and Poty 2016). The shale beds are separated by a sequence boundary, representing a sea-level fall between the transgressive events (Bond and Wignall 2008; Wendt 2021). Anoxic black shale associated with the Kellwasser Event is not recognised in all stratigraphic sections of the interval (e.g. Stephens and Sumner 2003), but the geographic extent of the black shale event confirms its global expression (Carmichael *et al.* 2019). This transgression-regression-transgression pattern is recorded in the type section of this interval at Coumiac in southern France (Bond and Wignall 2008), as well as many coeval sections — in the Rheinisches Schiefergebirge (Rhenish Mountains) of Germany (Bond and Wignall 2008), Poland (Racki *et al.* 2002), southern Belgium (Mottequin and Poty 2016), the Appalachian basin in New York State (ver Straeten *et al.* 2011; Beard *et al.* 2017),

the Canning basin of Australia (Stephens and Summer 2003), and in South China (Chen and Tucker 2003). In Morocco, this event is recognised as a single, major transgression and highstand that continues into the lower Middle Famennian (lower *rhomboidea* zone), persisting longer than in the type sections (France and Germany) (Lubeseder *et al.* 2010; Wendt 2021).

Sea-level change in the early Famennian is controversial. Some studies identified a major regressive surface in the basal Famennian; in Poland (Racki *et al.* 2002), in Alberta, Canada (Bond *et al.* 2013), in the Rheinisches Schiefergebirge (Schulke and Popp 2005), in the Canning Basin of Australia (Becker *et al.* 1991) in the Appalachian Basin (Pound Member of the Foreknobs Formation) (McClung *et al.* 2013; Brezinski *et al.* 2008), in Belgium (Dreesen and Thorez 1980) and in South China (Muechez *et al.* 1996; Zhang *et al.* 2019). Evidence for eustatic regression at this time from Belgium (Muechez *et al.* 1996) and Australia (Becker *et al.* 1991) is disputed. In the Belgian case, former conodont dating of the strata has been challenged (Mottequin and Poty 2016) and the Lower Famennian is considered transgressive. In the Australian Canning Basin, a basin-wide karstified surface at this boundary is alternately interpreted as a eustatic event (Becker *et al.* 1991) or a syn-sedimentary tectonic event (Chow *et al.* 2004).

Some research has, by contrast, indicated the lowermost Famennian to have been a transgressive interval (Königshof *et al.* 2016; Beard *et al.* 2017). In a broad review (and reinvestigation) of F–F boundary sections, including the stratotype section at Coumiac in France, Bond and Wignall (2008) found little evidence to support a eustatic regression in the lowermost Famennian. Rather, the uppermost Frasnian transgression was found to continue into the Famennian, and to be terminated by a sea-level fall in the Lower Famennian *triangularis* zone. Similarly, in the Moroccan Anti-Atlas, the lower Famennian recorded a highstand and the onset of lowstand deposition happened in the lower *rhomboidea* zone

(upper Lower Famennian) (Wendt 2021). For the North Devon Basin, the contact between the shallow-marine Morte Formation and the continental Pickwell Down Formation is considered to be Lower Famennian, indicating a sharp regression at this time, but age constraint on this contact is poor (Whittaker and Leveridge 2011).

A significant, widely recognised, two-pulsed regressive event occurred in the upper Lower Famennian (*rhomboidea* conodont zone), which is known as the Condroz Event (Königshof *et al.* 2016). This regressive interval is represented by the Esneux Formation in Belgium (Denayer *et al.* 2016) (Fig. 8.3D) and shallowing facies or hiatuses in Morocco (Lubeseder 2010; Wendt 2021) and South China (Zhang *et al.* 2019). Shallow water facies of the Nehden sandstone extended into basinal successions in the Rhenish Massif at this time (Late *rhomboidea* zone to Early *marginifera* zone) (Königshof *et al.* 2016) (Fig. 8.3A and B). The Condroz Event is thought to mark the beginning of an overall shallowing trend from the Middle to Upper Famennian, interrupted by several brief transgressions (Becker *et al.* 2012; Hartenfels and Becker 2016).

The Condroz regression is followed by a significant transgression in the lower *marginifera* zone, recognised in Morocco (Wendt 2021); Belgium (Johnson *et al.* 1986; Thorez *et al.*, 2006) and South China (Zhang *et al.* 2019), which is often referred to as the Enkeberg Event. The remainder of the Middle to late Famennian produced an overall shallowing trend, interrupted by brief transgressive pulses, including the *annulata* Event (upper *trachytera* zone) and the Dasberg Event (middle *expansa* zone) (Becker *et al.* 2020). The *annulata* Event is recognised by a pair of transgressive black shales in Poland (Racka *et al.* 2010; De Vleeschouwer *et al.* 2013) and Germany (Korn 2004; Myrow *et al.* 2011), in basinal sections in Morocco (Wendt 2021), and potentially by the Bon Mariage shale in Belgium (Thorez *et al.* 2006). Regression continued in the late Famennian (*postera* to Early *expansa* conodont zones), producing the most landward facies in Belgium with the Evieux Formation redbeds

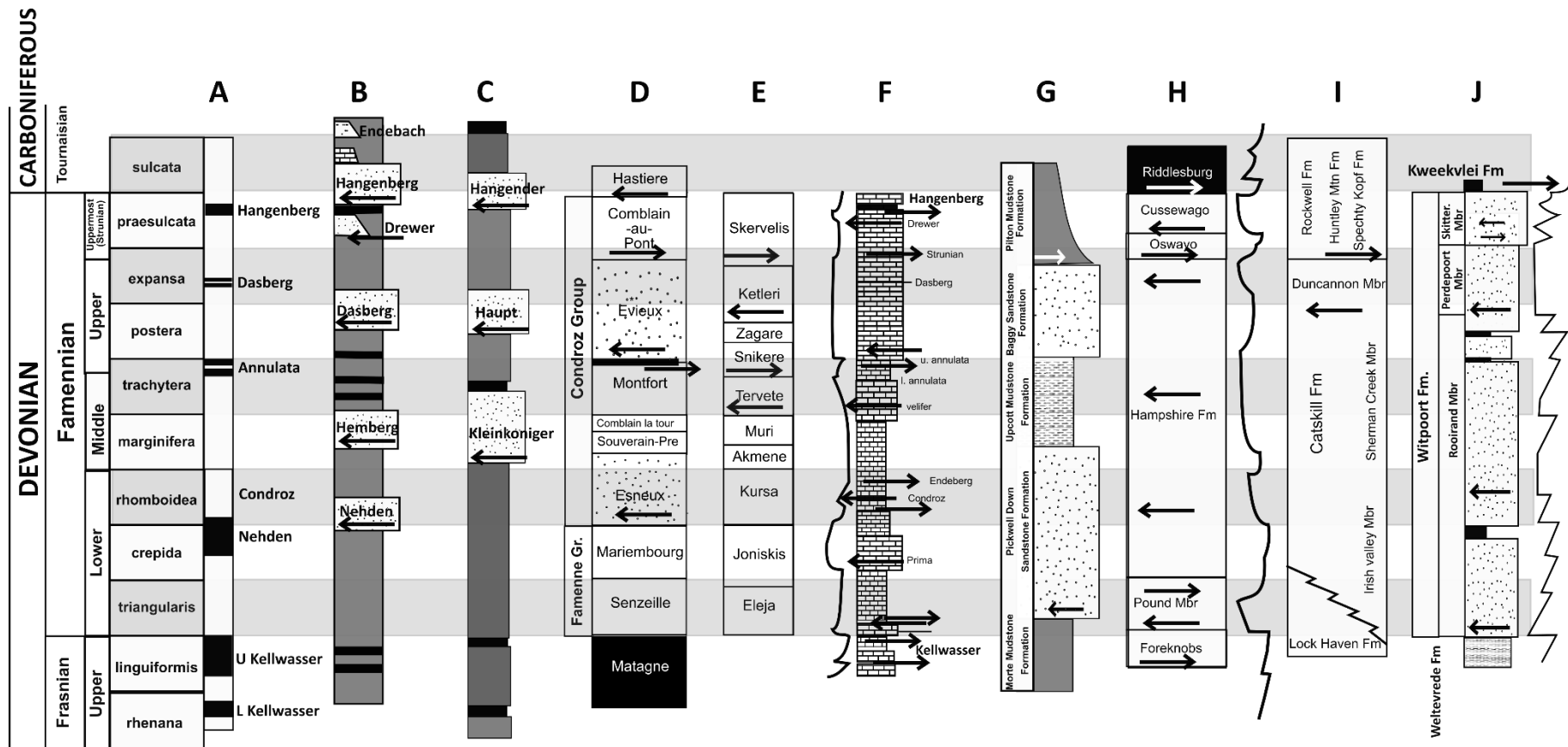
(Denayer *et al.* 2016), an overall shallowing trend in the Moroccan Anti-Atlas (Lubeseder *et al.* 2010), and significant hiatuses in SE Morocco (Hartenfels and Becker 2009), the maximum progradation of the Catskill clastic wedge in New York and Pennsylvania, represented by the Duncannon Member of the Catskill Formation (Elliot 2000), the Dasberg sandstones in basinal facies of the Rhenish Massif (Königshof *et al.* 2016), the Ketleri Formation of Latvia (Luksevics 2020), regressive trends in basinal sections in south China (Zhang *et al.* 2019), and is thought to account for the Acadian unconformity that underlies the Bakken Formation of the Williston Basin (Angulo and Buatois 2012). This corresponds to the top of T–R cycle IIe of Johnson *et al.* (1985). A brief transgressive Dasberg Event brought anoxic facies into many basins globally (Hartenfels and Becker 2009).

The latest Famennian (or Strunian) begins with a global sea-level rise in the Upper *expansa* zone. Progressive flooding is marked by the re-appearance of marine influence and recolonisation by marine infauna in many continental successions. The alluvial plain of the Catskill and Hampshire Formations in the Eastern USA was drowned by ‘transition beds’ that show the beginnings of marine transgression, including the Oswayo Formation, the Cleveland Shale and the Huntley Mountain Formation, which are all late to latest Famennian in age (Brezinski *et al.* 2009; Etensohn *et al.* 2023). The continental Old Red Sandstone facies was replaced by the marine-influenced Toe Head Sandstone Formation in the Irish Munster Basin (Higgs and Higgs 2015), the Pilton Mudstone Formation in the North Devon Basin (Whittaker and Leveridge 2011) and the Comblain-au-Pont Formation in Belgium, which transgressed the mostly non-marine Evieux Formation (Thorez *et al.* 2006; Denayer *et al.* 2021). This latest Famennian stepwise transgression continued into the Carboniferous Period, but was interrupted by rapid glacio-eustatic fluctuations with the Hangenberg Crisis in the Devonian–Carboniferous Boundary interval. A possible eustatic shallowing preceding the Hangenberg Event is represented by the Drewer Sandstone in the Rhenish Massif (Becker *et*

al. 2021), correlated with regressive facies in South China (Zhang *et al.* 2019), France (Feist *et al.* 2021) and Morocco (Kaiser *et al.* 2011). The Hangenberg Crisis is marked by a transgressive-regressive couplet of the Hangenberg Black Shale and overlying Hangenberg Sandstone of the Rhenish Massif (Germany) and their correlative strata (Lakin *et al.* 2016; Becker *et al.* 2016). A transgressive, anoxic black shale is recorded in many basins globally (Carmichael *et al.* 2016; Becker *et al.* 2016). The Hangenberg Sandstone of the Rhenish Mountains and its correlatives are considered to represent a glacio-eustatic regression (Becker *et al.* 2020). Correlatives of the Hangenberg Sandstone are recorded in the upper member of the Skervelis Formation in Latvia (Fig. 8.3E), which evidences long-lasting subaerial exposure at the DCB (Pipir *et al.* 2023), the basal bed of the Hastière Formation in Belgium (Denayer *et al.* 2021), the Middle Bakken Formation in the Western USA (Hogancamp and Pocknall 2018) and major erosional surfaces in Morocco (Kaiser *et al.* 2011; Wendt 2021). The Hangenberg Sandstone and its correlatives are thought to coincide with latest Famennian glacial deposits recognised in South America, the eastern United States and possibly Central and South Africa (Kaiser *et al.* 2011; Lakin *et al.* 2016; Lakin *et al.* 2021). Relative sea-level drop associated with the Hangenberg Sandstone event is estimated to have been as much as 100 m based on erosional incision in Morocco (Kaiser *et al.* 2011) and incised valleys in Europe (van Steenwinkel 1993; Bless *et al.* 1993) and Bolivia (Lakin *et al.* 2021). However, such a large-scale base-level fall is not evident in all successions (Denayer *et al.* 2021), and some sections, such as the Lali Section in South China (Zhang *et al.* 2019) and the Munster Basin of Ireland (Graham and Sevastopulo 2021), have no evidence of a regressive event at this time.

The uppermost Famennian to Tournaisian transgression continued following the Hangenberg Sandstone regression. An earliest Carboniferous deepening event is represented by the Riddlesberg Shale in the Appalachian Basin (Fig. 8.3H), the Castle Slate Formation in

the Munster Basin of Ireland (Higgs and Higgs 2015; Graham and Sevastopulo 2021), the Kasa Formation of Bolivia (Lakin *et al.* 2021), and the Stockum Limestone in the Rhenish Mountains (Becker *et al.* 2021). In the North Devon Basin (Fig. 8.3G), transgression began in the latest Famennian and continued into the Carboniferous, represented by the Pilton Mudstone Formation (Whittaker and Leveridge 2011).



LEGEND

→ **Transgression**

← Regression

 Shale/siltstone

Black shale


Carbonate
Sandstone Sandstone and shale

Unspecified

A - Famennian biostratigraphy and events (Koningshof et al. 2016)

B - Eastern Rhenish Succession, Germany (Koltonik et al. 2018)

C - Thuringia, Germany (Cronier 2007)

D - Belgium, Dinant syncinorium (Thorez et al. 2006)

E - Baltic Basin, Latvia (Luklsevics 2020)

F - South China, Lali Section (Zhang et al. 2019)

G - North Devon Basin (Whittaker and Leveridge 2011)

H - Appalachian Basin, Maryland (Brezinski et al. 2009)

I - Appalachian Basin, Pennsylvania (Brezinski et al. 2009)

J- Cape Basin, South Africa, (this work)

Figure 8.3 (previous page): Calibrated stratigraphic sections of Famennian strata with event stratigraphy. Only sections dated with conodont biostratigraphy are shaded with the biozones. Relative sea-level curves are given for the Appalachian Basin, South China and the Witpoort Formation.

8.6. Global correlatability of the Witpoort Formation

Cape Supergroup sedimentation patterns are often thought to be controlled by variations in the rate of subsidence (Tankard and Barwis 1982; Theron and Looock 1988; Hiller and Taylor 1992; Thamm and Johnson 2006; Tankard *et al.* 2012). The alternative hypothesis that eustatic events were the main control on the stratigraphy at the formational level (Cooper 1986) is tested below with reference to the known sea-level changes of the Famennian.

The Witpoort Formation is broadly consistent with a Famennian eustatic regression (Cooper 1986), but shorter term, third-order eustatic events are hitherto unrecognised in the succession. The Witpoort Formation is a coastal plain succession, and is therefore likely to have been highly sensitive to sea-level variations compared with deep-marine successions (Bond *et al.* 2013). Thick accumulations of FA 1 fluvial strata in the succession (up to tens of metres thick) represent episodes of maximum progradation, such as in the middle interval of the Rooirand Member and in the Perdepoort Member (Fig. 8.3J). In contrast, well developed fluvio-estuarine cycles in the lower and upper Rooirand Member are indicative of relatively higher rates of accommodation generation.

Formational contacts in the study area, at the Weltevrede–Witpoort and Witpoort–Kweekvlei boundaries also represent important shifts in accommodation. The former is interpreted as a transition from estuarine to fluvio-estuarine deposits, representing a seaward shift in the coastline position, but it is presently uncertain whether this contact is sudden or diachronous across the study area. The abrupt onlap of shale deposition represented by the Kweekvlei Formation represents a basin-wide flooding event (Browning and Reid 2017). It has long been predicted, based on lithostratigraphic and palaeontological grounds, that the

Kweekvlei Formation represents an earliest Carboniferous transgression (Cooper 1986; Theron 1994; Gess 2016).

The transition between the Perdepoort Member, Skitterykloof Member and Kweekvlei Formation records a series of base-level changes that can be recognised globally towards the end of the Devonian Period. The exclusively fluvial strata of the Perdepoort Member record maximum regression in the Witpoort Formation. Similarly, maximum progradation during the late Famennian is represented by the Duncannon Member of the Catskill Formation (Slingerland *et al.* 2009; Broussard *et al.* 2018), the Evieux Formation in Belgium (Crupe Member) (Thorez *et al.* 2006), the Dasberg sandstone in the Rhenish successions (Königshof *et al.* 2016) and the Ketleri Formation in Latvia (Luksevics 2020). The Perdepoort Member is a likely correlative of all of these units, reflecting a late Famennian eustatic regression (Fig. 8.3). The Skitterykloof Member and its eastern correlative record the beginnings of eustatic transgression in the latest Famennian, as indicated by the re-appearance of estuarine mudrock. This Strunian transgression is recognised in many basins. The DCB interval in Bolivia, as described by Lakin *et al.* (2021) bears a strong resemblance to the upper Witpoort Formation (Fig. 8.4). The upper 100 m of the Colpacucho Formation is a sandy shoreface deposit that regressively overlies offshore marine shale and is overlain by the ‘Cumana Formation equivalent unit’, which is a conglomeratic or diamictite unit that is 58 to 140 m thick. A sharp transition to the offshore marine shale-dominated Kasa Formation caps the Famennian succession. The Devonian–Carboniferous Boundary is placed at 12 m above the flooding surface at the base of the Kasa Formation, based on the last appearance of the latest Famennian index miospore *Retispora lepidophyta* (it should be noted, however, that the terrestrial extinctions occurred slightly later than the marine expression of the Hangenberg Event). Equivalence between South African and Bolivian Famennian strata was previously suggested by Almond *et al.* (2002). The regressive Perdepoort Member, gravelly

Skitterykloof Member and shaly Kweekvlei Formation present an analogous succession to that of Bolivia, and reflect the eustatic events of this interval globally (Fig. 8.4). Lithic gravel beds that occur in the Skitterykloof Member, which are unknown in stratigraphically lower strata in the Devonian of South Africa, may be explained by glacial transport of lithic clasts into the depositional basin. A possible expression of the transgression–regression couplet of the Hangenberg Black Shale and the Hangenberg Sandstone, is found in all three studied sections of the Skitterykloof Member (Fig. 8.5), in which a 1–2 m thick bioturbated shale or siltstone bed is sharply overlain by a conglomeratic horizon. In sections WP 2 and WP 3, this horizon is approximately 10 m below the Witpoort–Kweekvlei contact, whereas in section AD the horizon is approximately 5 m below the contact.

Based on the above evaluations, it is expected that other Famennian regressive and transgressive events should be reflected in the underlying Rooirand Member strata. Immediately underlying the Perdepoort Member, a conspicuous shale-rich interval (Waterloo Farm sections and lateral equivalents) is consistent with the transgressive *annulata* Events that occurred at the Middle–Upper Famennian boundary (*trachytera* conodont biozone). A regressive interval in the middle Rooirand Member may reflect the eustatic Condroz Event regression of the Lower-Middle Famennian (Fig. 8.3), which would in turn place the regressive contact at the base of the formation in the Lower Famennian. This Lower Famennian sea-level fall is controversial; as reviewed above, some studies infer a sea-level fall at the F–F boundary, whereas others recognised regression later in the Lower Famennian (Fig. 8.3). Moreover, as is evident in the Catskill Formation and equivalent strata in the Appalachian Basin (Broussard *et al.* 2020), regressive contacts are often diachronous and do not necessarily record a precise datum. This analysis predicts that the Kellwasser Event would be recorded in the upper Weltevrede Formation, representing a global Late Frasnian transgressive interval. A better understanding of sequence stratigraphic patterns in this

formation, and a search for geochemical, sedimentological and palaeontological evidence, would aid the recognition of this event.

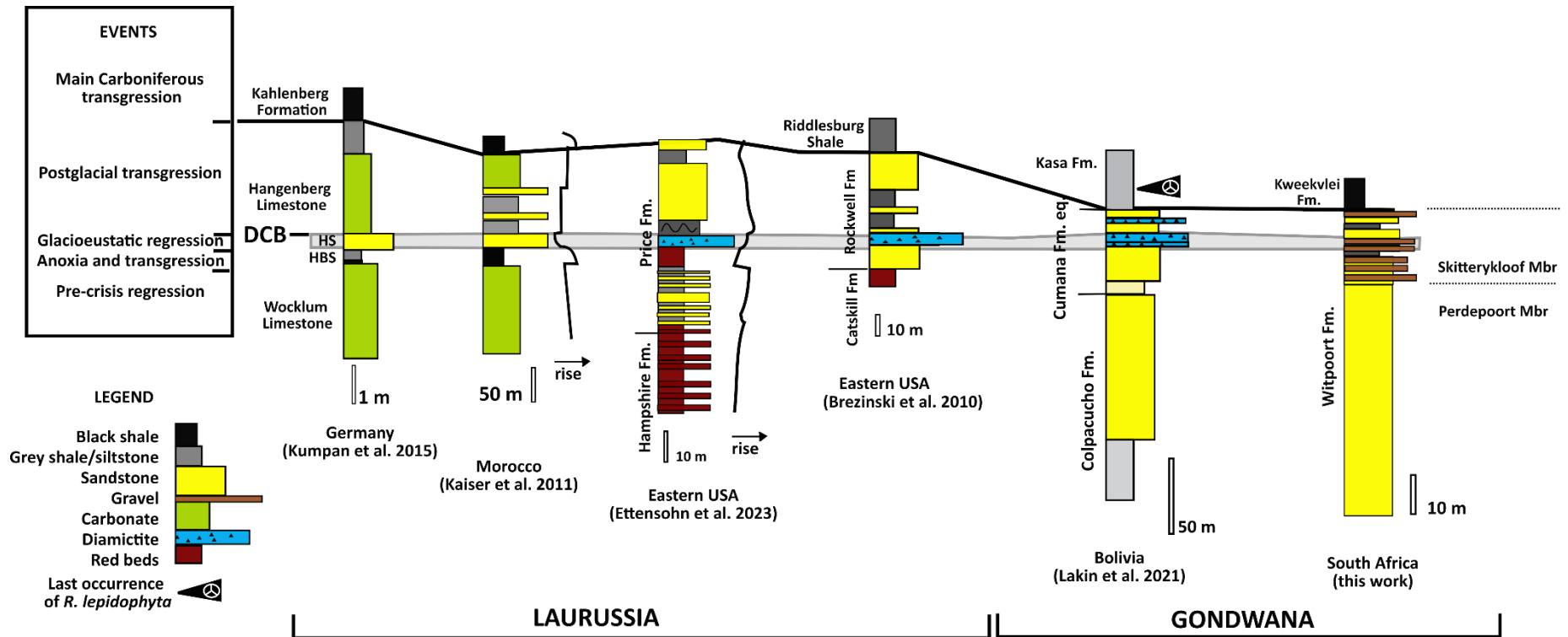


Figure 8.4: Devonian–Carboniferous Boundary interval in various stratigraphic sections worldwide, correlated according to event stratigraphy. The Devonian–Carboniferous Boundary (DCB) is placed at the top of the Hangenberg Sandstone and its correlatives, according to recent proposals by the International Working Group on the DCB. HBS = Hangenberg Black Shale, HS = Hangenberg Sandstone.

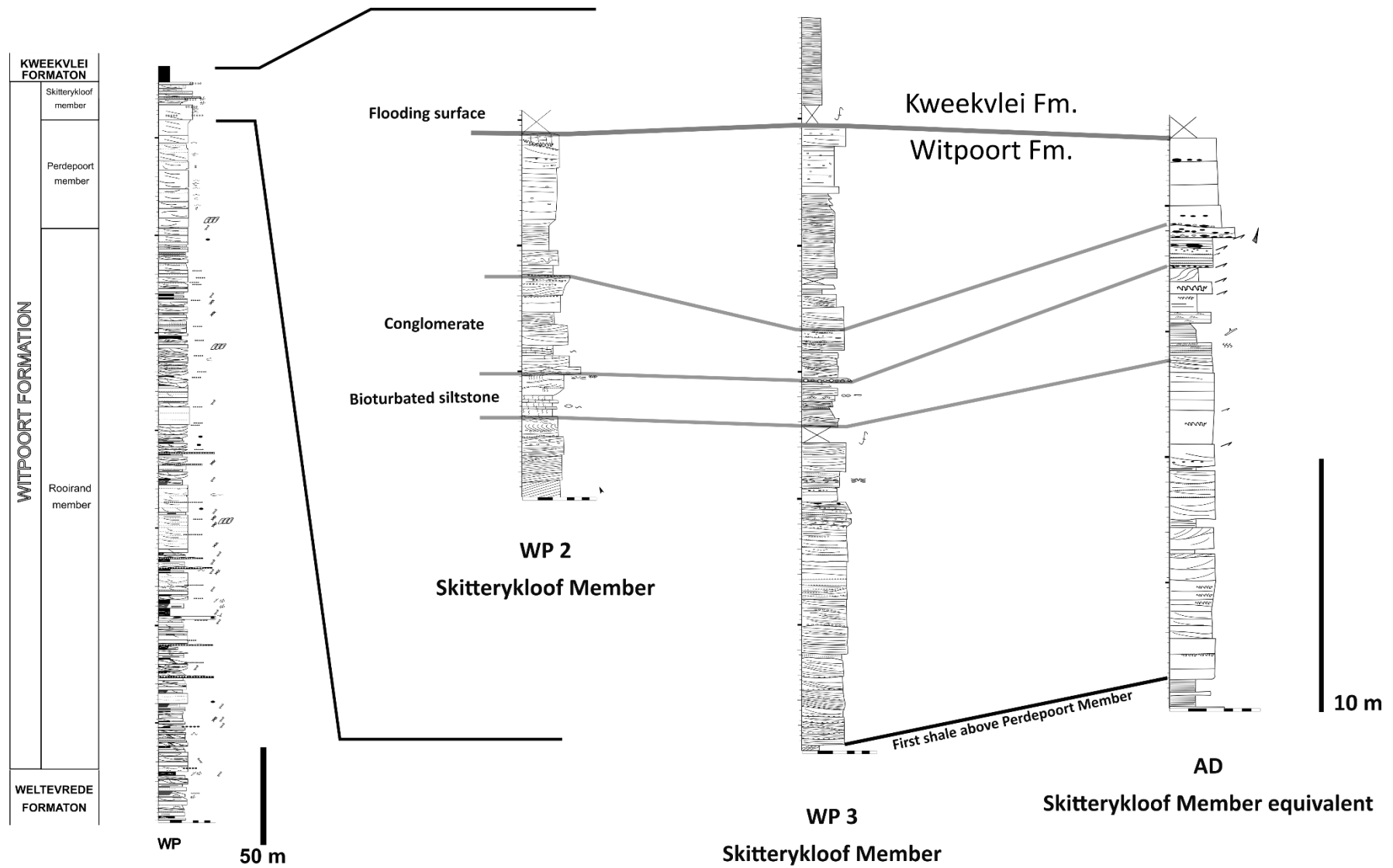


Figure 8.5: The Devonian–Carboniferous Boundary interval in South Africa, recorded in the Skitterykloof Member at the top of the Witpoort Formation, as illustrated by three stratigraphic sections from the study area.

Given that global third-order eustatic patterns appear to be recorded in the strata of the Witpoort Formation, it is also possible that higher frequency eustatic cycles may also have influenced the stratigraphic succession. The question of short-term sea-level change in the Famennian is complicated because, although high-frequency Milankovitch cyclicity is evident in Famennian basinal deposits (De Vleeschouwer *et al.* 2014), it remains uncertain whether this reflects sea-level or climatic changes, though it could be expected that the two are interrelated phenomena.

High-frequency sedimentary cyclicity, forming metre- to tens-of-metre scale fluvio-estuarine cyclothems in the Rooirand Member, indicates repeated base-level changes. Plotting this high-frequency cyclicity to the Witpoort Formation stratigraphic column produces 18 middle basin and 22 eastern sector (composite section) muddier-upward cycles for the Rooirand Member (Fig. 8.6). This plot indicates an order of cyclicity that is consistent with that known in other Famennian successions. For example, studies of marine shelf deposits have recognised cyclical variations in magnetic susceptibility (MS) that are considered to reflect 100 kyr and 405 kyr orbital periods during the Famennian, related to the eccentricity of the Earth's orbit (De Vleeschouwer *et al.* 2013; Pas *et al.* 2018; Ma *et al.* 2022). In the Illinois Basin, for example, 31 eccentricity cycles (405 kyr) are projected for the Famennian interval (Pas *et al.* 2018), somewhat similarly to the Lali Section of South China, where 37 cycles were identified for the interval (Ma *et al.* 2022). This 405 kyr orbital period is thought to have exerted a strong control on repetitive facies changes in basinal deposits in Poland, where nine similar cycles occur in the Lower Famennian succession at Kowala quarry (De Vleeschouwer *et al.* 2013). The fluctuations in MS are thought to indicate shifts in

detrital input to marine basins related to climate change, and possibly eustasy (Whalen and Day 2010; De Vleeschouwer *et al.* 2013, 2014; Pas *et al.* 2018). If cyclicity in the Witpoort Formation is taken to reflect fourth-order (0.1–0.5 Myr), eccentricity-driven base-level change, the 20 or so sequences in the Rooirand Member would therefore reflect much of Famennian time (Fig. 8.6). This inference is made with the utmost tentativeness because the sequence stratigraphy of non-marine or paralic strata is far more complex than in continuously deposited, deep marine strata. Cyclicity in the Rooirand Member will likely have been affected by hiatus and erosion at sequence boundaries that could encompass the falling stages of a base-level cycle, coupled with far more rapid accumulation of sediment during the rising stages. Incompleteness in the stratigraphic record implies that direct correlations of Rooirand Member cyclothems to a Famennian orbital period is, therefore, probably unlikely based on present evidence. The strata nonetheless appear to show similar orders of base-level change to the known Famennian eccentricity-driven orbital period. It is, furthermore, speculated in Figure 8.6 that fluvially-dominated intervals in the Witpoort Formation stratigraphic column (Fig. 8.6B and C) appear to correlate with intervals of low eccentricity, inferred from the floating astronomical time scale for the Famennian given by Pas *et al.* (2018) (Fig. 8.6A). Fluvial progradation would be expected during such intervals because low eccentricity has been linked to increased continental glaciation (De Vleeschouwer *et al.* 2024). High eccentricity intervals, on the other hand, may have led to ice sheet collapse forming the transgressive anoxic events that punctuate the Famennian (De Vleeschouwer *et al.* 2013). If verified, this correlation would make a strong case for an allogenic control on base level in the Witpoort Formation. The result would be consistent with inferred base-level change in the Lower Famennian strata of the upper Foreknobs Formation (McClung *et al.* 2013), in which 20–100 m thick, upwards-shallowing, shallow-marine parasequence sets are thought to record ~300 kyr fluctuations in relative sea level. Far

thinner cycles in the Witpoort Formation could represent similar fourth-order base-level change because rate of accommodation would have been far lower in the Cape Basin than in the Appalachian Basin at this time.

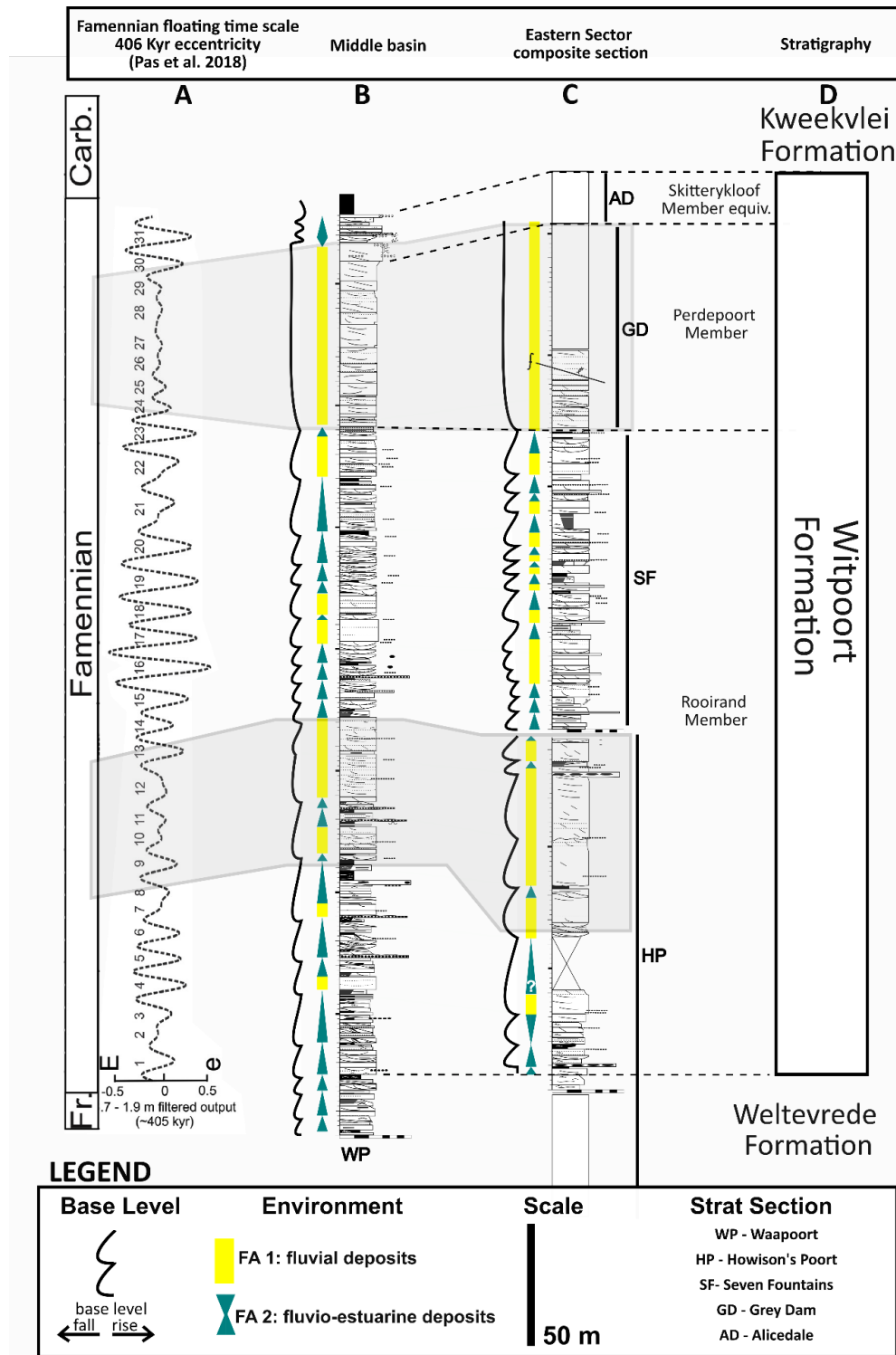


Figure 8.6: Fourth-order cyclicity for the Famennian, which appears to be reflected in the Witpoort Formation stratigraphic column. A) Eccentricity-driven 405 kyr cycles during the Famennian based on magnetic susceptibility. B–C) Composite stratigraphic columns for the Witpoort Formation in the study area with interpretations of small scale base-level changes. Grey areas indicate intervals of eccentricity minima in A, which appear to correspond to weakly cyclical, fluvially-dominated portions of the stratigraphic succession in B and C.

8.7. Palaeoecology

The ecology and environments of the Witpoort Formation (and wider Witteberg Group) are paramount to understanding the global expression of the Late Devonian biotic crisis. Exceptional fossil preservation at Waterloo Farm, and the intensive sampling thereof, provides one of the most diverse floral and faunal ecosystem records of the Famennian (Gess and Whitfield 2020). For this time stage, Waterloo Farm is the only high-latitude palaeoecological dataset, apart from preliminary palaeobotanical studies on fossiliferous horizons at Howison's Poort (Plumstead 1967; Anderson and Anderson 1985) and Coombs Hill (Harris *et al.* 2021a, b). A new fossil locality from Rabbit Ridge, introduced in Chapter 7, contributes another locus for understanding the biota of this poorly explored palaeo-region.

The known invertebrate ecology and plant taphonomy of the formation was explored in Chapter 7 (7.6). The Rabbit Ridge locality preserves the youngest marine invertebrates known from the Cape Supergroup. Their superabundance in that deposit is interpreted as indicating a more distal, and marine-influenced, lagoonal setting than is apparent from Waterloo Farm. This is consistent with trace fossil diversity and plant taphonomy from the lingulid brachiopod-bearing horizon. A wider analysis of the sedimentary facies in the Rooirand Member, comprising the lower $\frac{3}{4}$ of the formation's thickness, has indicated repetitive fluvio-estuarine cyclothems. Metres-thick accumulations of mudrock, such as that described from Rabbit Ridge and Waterloo Farm, are therefore expected to have developed in wave-dominated estuary basins, comprising fresh- to brackish-water conditions. It was hypothesised in Chapter 7 that salinity fluctuations in such a system could have contributed to mass mortalities of lingulid brachiopods, though inadequate attention was given to the rate of deposition, which in estuaries can reach metres *per annum*. An equally plausible kill mechanism for lingulid brachiopods is their limited escape potential due to anchorage of the pedicle at the base of the burrow. Rapid sedimentation may result in detachment from the

anchored pedicle resulting in the exhumation of the brachiopod from its burrow, or interment of the animal (Hammond 1983).

The Waterloo Farm Main Fish Lens (MFL) represents fresh- to brackish-water conditions, as evidenced by the fauna and flora contained therein (Gess and Whitfield 2020). This assessment gains support from ichnotaxonomic study of the locality (6.3). The present analysis indicates that the accumulation of black shale occurred in the estuary turbidity maximum zone, which is likely to have occurred at some distance from the shoreline. The turbidity maximum in modern estuaries may extend for tens of kilometres in a shoreline-perpendicular direction and may reach as far as the salinity limit, or further upstream (Sandford *et al.* 2001; Uncles *et al.* 2006). It is entirely plausible therefore that the deposit occurred under partially freshwater conditions. The extent of the salinity limit landwards from the shoreline is controlled by interplays of fluvial discharge, tidal range and coastal gradient, among other factors (Van den Berg *et al.* 2007). Episodes of low river discharge enable greater incursions of tidal marine waters, and, given the strong seasonality expected for this high-latitude region and the evident flashy discharge inferred from fluvial facies in the formation (FA 1), the presence of brackish water in the immediate environment of the MFL may have been ephemeral. This effect would, hypothetically, have been exacerbated by seasonal influx of meltwaters from mountain glaciers in the source area (Gess and Whitfield 2020).

In this work that the possibility of ephemerally frozen landscapes in the Witpoort Formation is proposed, as is indicated by the occurrences of dropstones and a sharp-walled clastic dyke at Waterloo Farm, and transported, probably frozen, sand clasts at various horizons in the stratigraphic package. Possible ice keel scour marks in the uppermost Rooirand Member, and the appearance of lithic clasts at the top of the formation provide further putative evidence of periglacial processes. This evidence, however, does not amount

to proof of glaciation in the late to latest Famennian in South Africa, which has already been suggested (Almond *et al.* 2002). It should also be noted though that neo-sedimentological studies on seasonally frozen braided river systems such as the Tana River in Norway and the Saskatchewan River in Canada have produced no evidence of ice-related sedimentary structures (Collinson 1970; Cant and Walker 1978).

Stratigraphic investigation of Waterloo Farm (Chapter 4) places the deposit in the upper Rooirand Member, significantly underlying the top of the Witpoort Formation. The locality is, therefore, likely older than prior, recent assessments have postulated. According to the event stratigraphy review above, Waterloo Farm predates a eustatic sea-level fall in the late Famennian that is correlated to the Perdepoort Member. It is, therefore, hypothesised that a eustatic transgression at the Middle–late Famennian boundary, the *annulata* Event (*trachytera* condont zone), potentially flooded the coastal plain during deposition of the MFL. It is, further, speculated that the two-pulsed *annulata* Event could explain the apparent paraconformity between the MFL and its laterally adjacent Blue Layer. This hypothesis puts the locality closer in age to late Famennian tetrapod-bearing localities in Laurussia such as Red Hill in eastern USA, Strud in Belgium, and the Pavāri in Latvia. Palynology places both Red Hill and Strud in the VCo miospore biozone (Cressler *et al.* 2010; Denayer *et al.* 2016), an approximate correlative of the *postera* conodont zone in the early late Famennian (Fig. 8.3).

Strud and Red Hill are both considered to represent non-marine fluvial floodplain deposits, with possible marine influence (Cressler *et al.* 2010; Denayer *et al.* 2016). In the case of Red Hill, the occurrence of a single ctenacanth shark spine in the deposit indicates a connection to open-marine environments, though the specimen is an outlier, representing a taxon that is common in contemporary open-marine deposits (Cressler *et al.* 2010).

Surprisingly, there appears to be a high degree of palaeoecological congruence between the three sites, given the great latitudinal differences between the tropical climes of Strud and Red Hill and the sub-polar region encompassing Waterloo Farm. This is evidenced by higher-order taxonomic diversity of vertebrates and plants, and presence of terrestrial invertebrates at Waterloo Farm, Strud and Red Hill.

Plant diversity was overall similar, with archaeopterid forests and zygopterid ferns as predominating taxa. The fern-like plant *Rhacophyton* is common at both Red Hill and Strud, whereas *Flabellopteris* is abundant at Waterloo Farm. The latter is a plant of uncertain affinity, which nonetheless shares some morphological features with the Rhacophytales (Gess and Prestianni 2022), and may have occupied similar ecological niches. *Archaeopteris* is common at all three sites, representing a globally distributed large tree that would have seasonally shed its ‘fronds’ (Gess and Berry 2024). Lycopods, including arborescent and herborial forms are prolific at both Red Hill and Waterloo Farm, but are not found at Strud. The paucity of lycopods in the Belgian Devonian strata has been explained by arid climates in this region of the Old Red Continent, whereas lycopods at Red Hill and other localities in the Catskill Mountains are indicated to have occupied wetland environments (Scheckler 1986; Fairon-Demaret 1996). Spermatophytes (seed plants) were present at all three localities; at Red Hill the seed plants are indicated to have opportunistically occupied habitats disturbed by fire (Cressler *et al.* 2010).

In terms of vertebrate animal diversity, the three sites mentioned all contain groenlandaspigid placoderms, acanthodians and various Sarcopterygii, including dipnoans, tristichopterids and tetrapods. Waterloo Farm is the only locality of the three containing a diversity of sharks as well as coelacanths and lampreys, which are indicative of marine influences at the site (Gess and Whitfield 2020). Among tristichopterids, the large predator *Hyneria* represents a genus only found at Waterloo Farm and Red Hill (Gess and Ahlberg 2023). Placoderms from Red

Hill and Waterloo Farm share remarkable similarity, with *Groenlandaspis* occurring at both sites and the smaller groenlandaspid *Africanaspis* at Waterloo Farm being similar, and perhaps congeneric to, *Turrisaspis* from Red Hill and Strud (R. W. Gess, pers. comm.). The gyracanthid acanthodian *Gyracanthides riniensis* from Waterloo Farm bears a strong resemblance in spine ornamentation to a gyracanthid from Red Hill, '*Gyracanthus*' *sherwoodi* (Gess and Burrow 2024). Phyllolepid placoderms are abundant at Strud and Red Hill, though absent from the Waterloo fauna. The absence of cosmopolitan antiarch *Bothriolepis* at Red Hill is unusual because the genus is widely distributed in the Catskill Formation, and other non-marine to marginal-marine sites globally in the Famennian (Cressler *et al.* 2010), including Waterloo Farm and Strud. A further point of similarity between the Catskill Formation and the Witpoort Formation is the presence of bivalves in great abundance, including in apparent life position. There are nuances in biodiversity and variations between the sites that require careful analysis to understand intercontinental biodiversity patterns in the Famennian, which is beyond the scope of the present overview. Overall, the biodiversity supports a notion of a high degree of cosmopolitanism in vertebrates and plants during the Middle to late Famennian. Comparability between Waterloo Farm and geographically distant, non-marine (possibly slightly marine-influenced) localities is consistent with the sedimentological and ichnological evidence from Waterloo Farm, pointing to a fluvio-estuarine sedimentary system with fluctuations in river discharge controlling the interplay of freshwater and brackish water environments.

8.8. Conclusions

1. The Witpoort Formation in the Eastern Cape is a non-marine to marginal-marine deposit that accumulated along a WNW-ESE trending Agulhas Sea shoreline. Though connections to open-marine environments are indicated by body and trace fossils from

the marginal-marine strata, open-marine deposits are not recognised in the preserved succession.

2. The three members of the Witpoort Formation, the Rooirand Member, the Perdepoort Member and the Skitterykloof Member (or an equivalent thereof) are recognised across the study area for the first time.
3. Rooirand Member strata accumulated in retrogradational, erosionally-bound cyclothems that are consistent with incised-valley fill facies models, whereas the Perdepoort Member is a fluvial deposit. The Skitterykloof Member represents a transition zone from the arenaceous Witpoort Formation to the overlying argillaceous Kweekvlei Formation with the introduction of erratic lithic pebbles possibly indicating glaciogenic transport into the basin.
4. The hypothesis (Cooper 1986) that the Witpoort Formation represents a Famennian eustatic regression is consistent with the evidence presented in this thesis, indicating the maximum regression of the Devonian system in South Africa. Higher-frequency third order, and possibly fourth order, eustatic shifts are also recognised in the succession. It is hypothesised that global sea-level change exerted a strong control on sedimentation in the Witpoort Formation at various orders of frequency.
5. The stratigraphic framework posited here places the Waterloo Farm fossil locality as Middle to early late Famennian, rather than latest Famennian. This result has implications for patterns of intercontinental faunal and floral exchange during the Famennian. Further constraint on the age of this and other localities in the Witpoort Formation should be attempted with continued stratigraphic, biostratigraphic, geochemical and micropalaeontological (especially palynological) research in the Witteberg Group.

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APPENDICES

Appendix 1A – GPS locations of stratigraphic sections

Locality	Strat section	Base	Top
Rabbit Ridge	RR	33°17'57.71"S 26°45'9.68"E	33°17'47.81"S 26°44'33.63"E
Waterloo Farm	WF 1a	33°19'24.88"S 26°32'19.21"E	33°19'21.49"S 26°32'3.19"E
	WF 1b	33°19'22.56"S 26°32'13.20"E	33°19'11.69"S 26°32'17.72"E
	WF 2	33°19'20.42"S 26°32'0.38"E	33°19'18.95"S 26°31'44.58"E
Grey Dam	GD	33°19'20.41"S 26°31'29.87"E	33°19'23.72"S 26°31'18.65"E
Howison's Poort	HP 1	33°22'19.84"S 26°28'30.14"E	33°22'18.14"S 26°28'31.48"E
	HP 2	33°22'15.81"S 26°28'33.37"E	33°22'13.95"S 26°28'34.61"E
	HP 3	33°22'10.69"S 26°28'35.75"E	33°22'6.88"S 26°28'36.80"E
	HP 4	33°22'8.41"S 26°28'40.12"E	33°22'7.22"S 26°28'42.38"E
Seven Fountains	SF	33°27'16.31"S 26°16'51.39"E	33°27'10.53"S 26°16'57.81"E
Alicedale	AD	33°19'46.67"S 26° 4'39.57"E	33°19'45.52"S 26° 4'40.27"E
Waaipoort	WP 1	33°14'54.38"S 24°29'40.88"E	33°14'46.22"S 24°29'48.12"E
	WP 2	33°14'37.16"S 24°29'50.07"E	33°14'36.20"S 24°30'1.44"E
	WP 3	33°14'42.13"S 24°29'49.25"E	33°14'40.61"S 24°29'50.65"E

Appendix 1B – Legend to stratigraphic log sections

SEDIMENTARY FACIES

	Sm - Structureless sandstone
	Sl (p) Low-angled planar cross-stratification
	SI - Low-angled trough cross-stratification
	St - Trough cross-stratification
	Sp - Planar cross-stratification
	Sh - horizontal lamination
	Su - undulatory lamination
	Fh - horizontally laminated shale/mudstone

SEDIMENTARY STRUCTURES

	Sand clast
	Mudrock clast
	Discontinuous mud clast stringer
	Continuous mud clast stringer
	Lithic clast stringer
	Ripple form sets
	Clay drape
	Ice keel scour marks (?)
	Adhesion ripples
	Heavy mineral laminae
	Parabolic cross-stratification
	Chevrons
	Convolution
	Unidirectional ripples
	Combined flow ripples
	Clay-draped ripples
	Gutter
	Ferruginous concretion

STRUCTURAL GEOLOGY

	Large-scale fault
	Small-scale fault
	Brecciation
	Shearing in mudrock
	Joints
	Tension gashes
	Slickenside

RESTORED FLOW VECTORS

	flow vector
	trough axis orientation
	flow trend

STRATIGRAPHIC CONTACTS

	Kweekvlei Fm.
	Witpoort Fm.
	Weltevrede Fm.
SK	Skitterykloof Member
PP	Perdepoort Member
RR	Rooirand Member

ICHNOTAXA

<i>Altichnus</i> spp.	
<i>Archaeonassa fossilata</i>	
<i>Arenicolites curvatus</i>	
<i>Cochlichnus anguineus</i>	
<i>Cruziana problematica</i>	
<i>Gordia marina</i>	
<i>Lingulichnus verticalis</i>	
<i>Lockeia siliquaria</i>	
<i>Nereites</i> spp.	
<i>Palaeophycus tubularis</i>	
<i>Phycodes</i> cf. <i>palmaris</i>	
<i>Planolites</i> spp.	
<i>Psammichnites implexus</i>	
<i>Ptychoplasma exelsum</i>	

BODY FOSSILS

	Plant axis
	Plant leaves
	Plant fructifications
	Comminuted plants
	Chondrichthyan spine
	Fish fossil
	Lingulid brachiopod

GRAIN SIZE

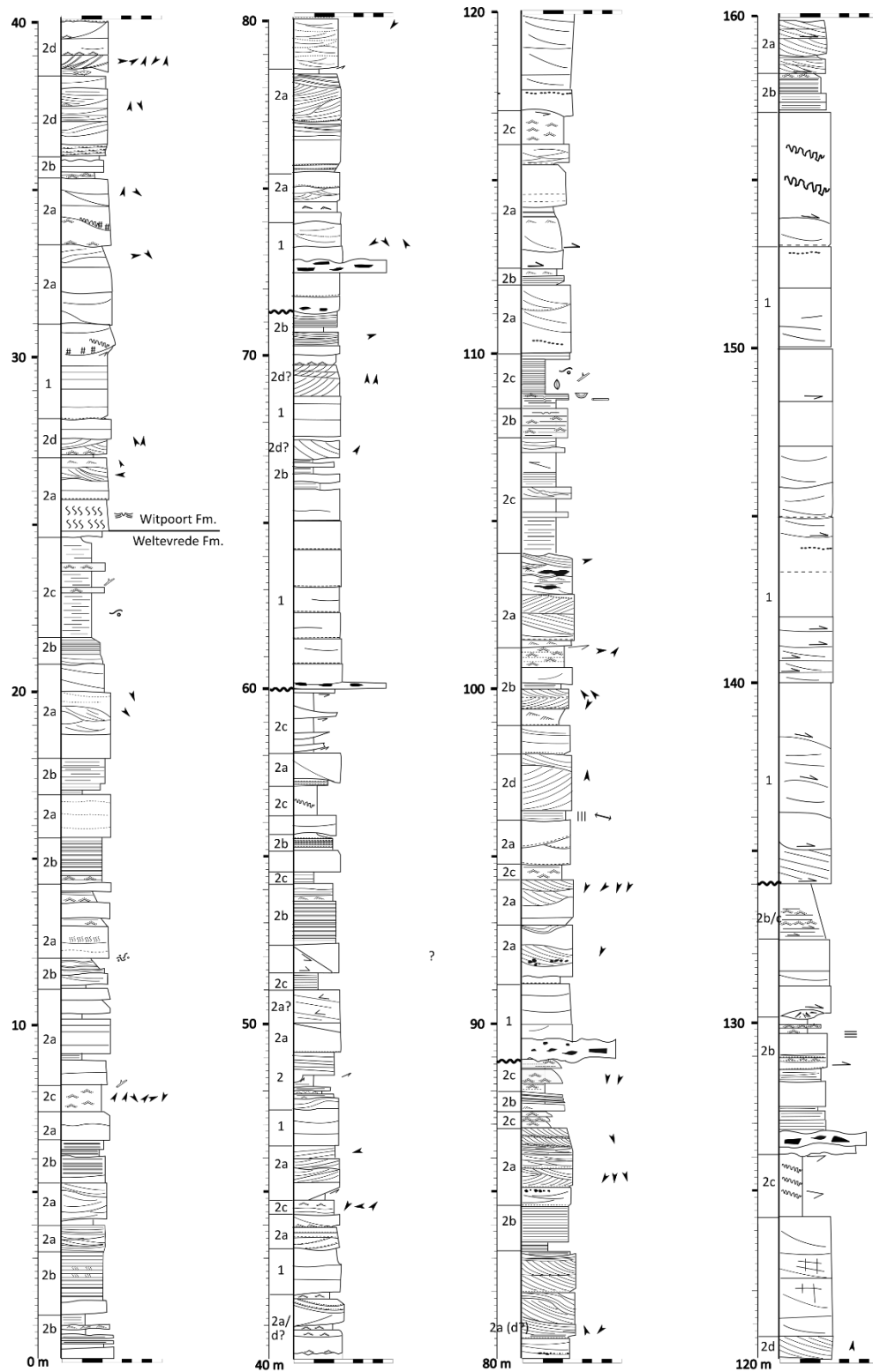
	clay
	silt
	sand
	granule
	pebble
	cobble
	boulder

FACIES ASSEMBLAGE

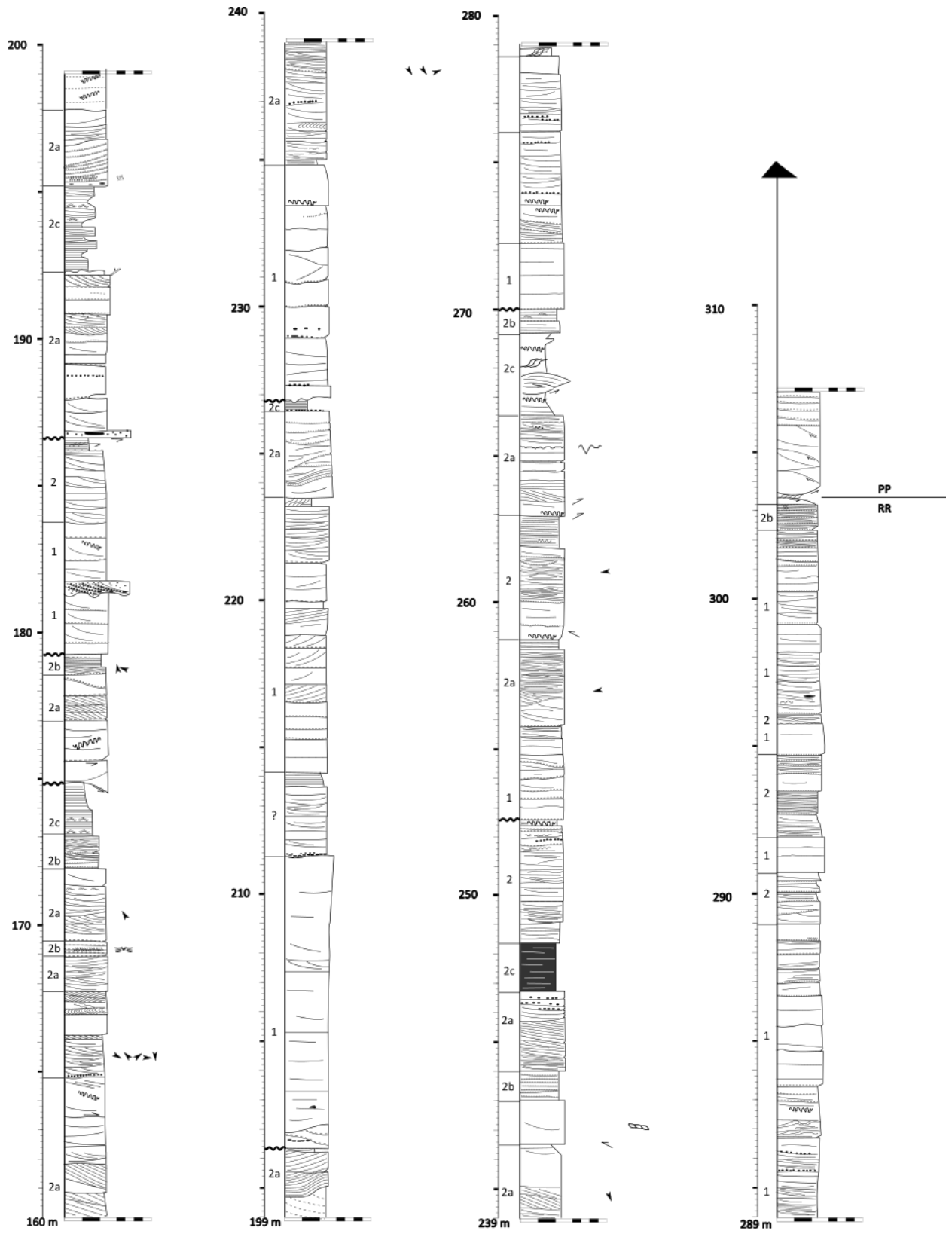
2c	Muddy estuary basin
2b	Estuary flats
2a	Estuary channels / Bayhead delta
1	Braided river channels and bars

<i>Ptychoplasma conica</i>	
<i>Rusophycus carbonarius</i>	
<i>Scalichnus lepusaures</i>	
<i>Selenichnites rossendalis</i>	
<i>Undichna</i> spp.	
<i>Zoophycos</i> cf. <i>brianteus</i>	
<i>Spirophyton</i>	
Plug-shaped burrows	
Compound trace 1	
Compound trace 2	
Bromalites	
Traces indet./bioturbation	
Diminutive burrows indet.	
Vertical burrows indet.	
Siphonichnidae indet.	

Appendix 1C – Section WP 1



WP 1 (1/2)

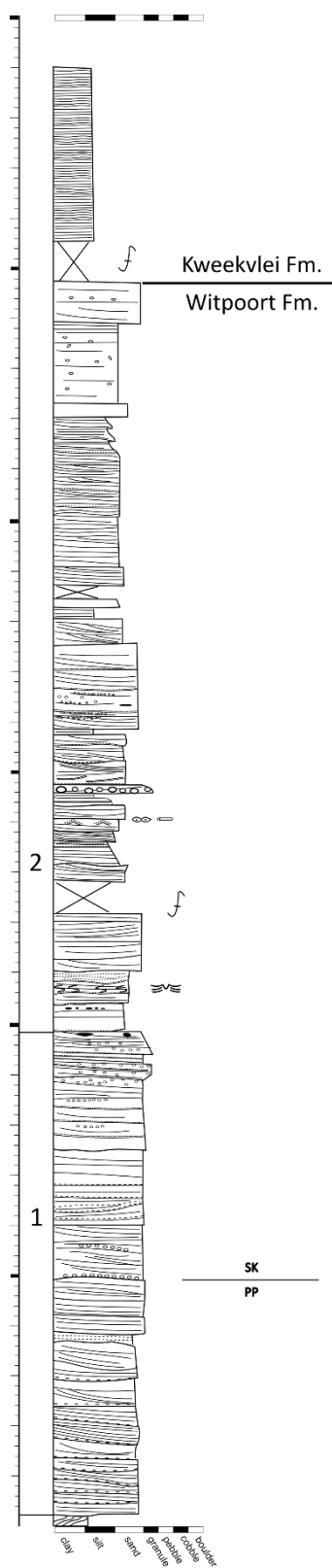


WP 1 (2/2)

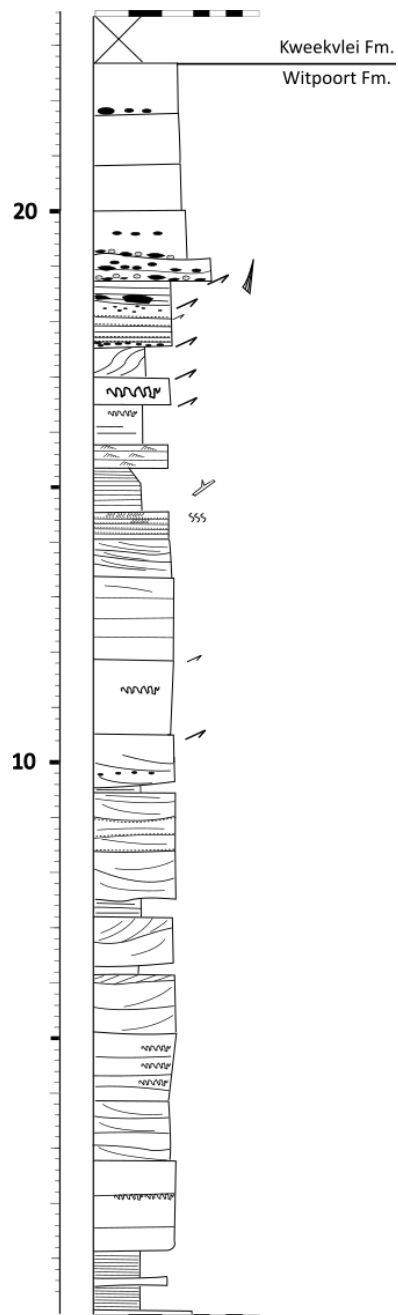
Appendix 1D – Section WP 2



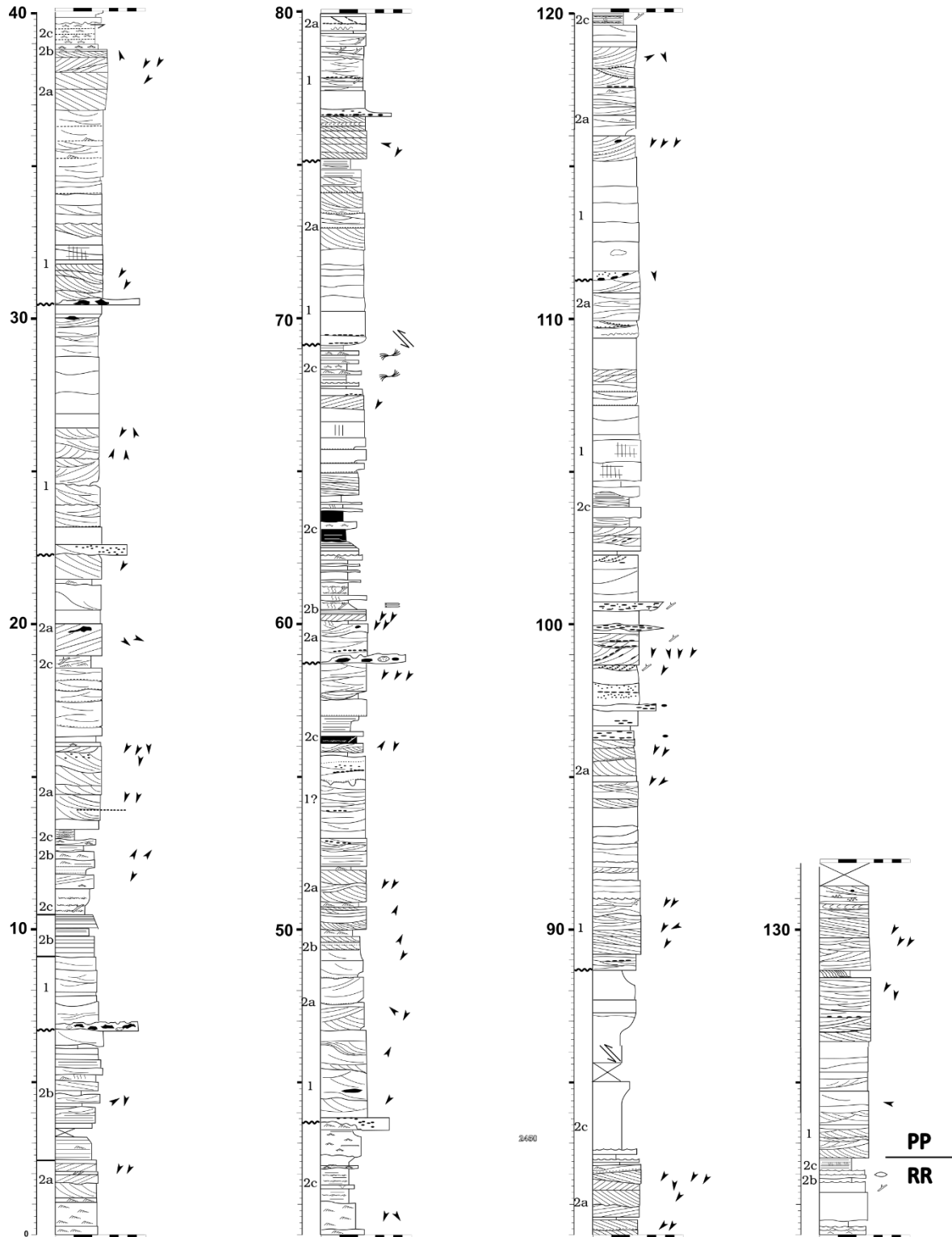
Appendix 1E – Section WP 3



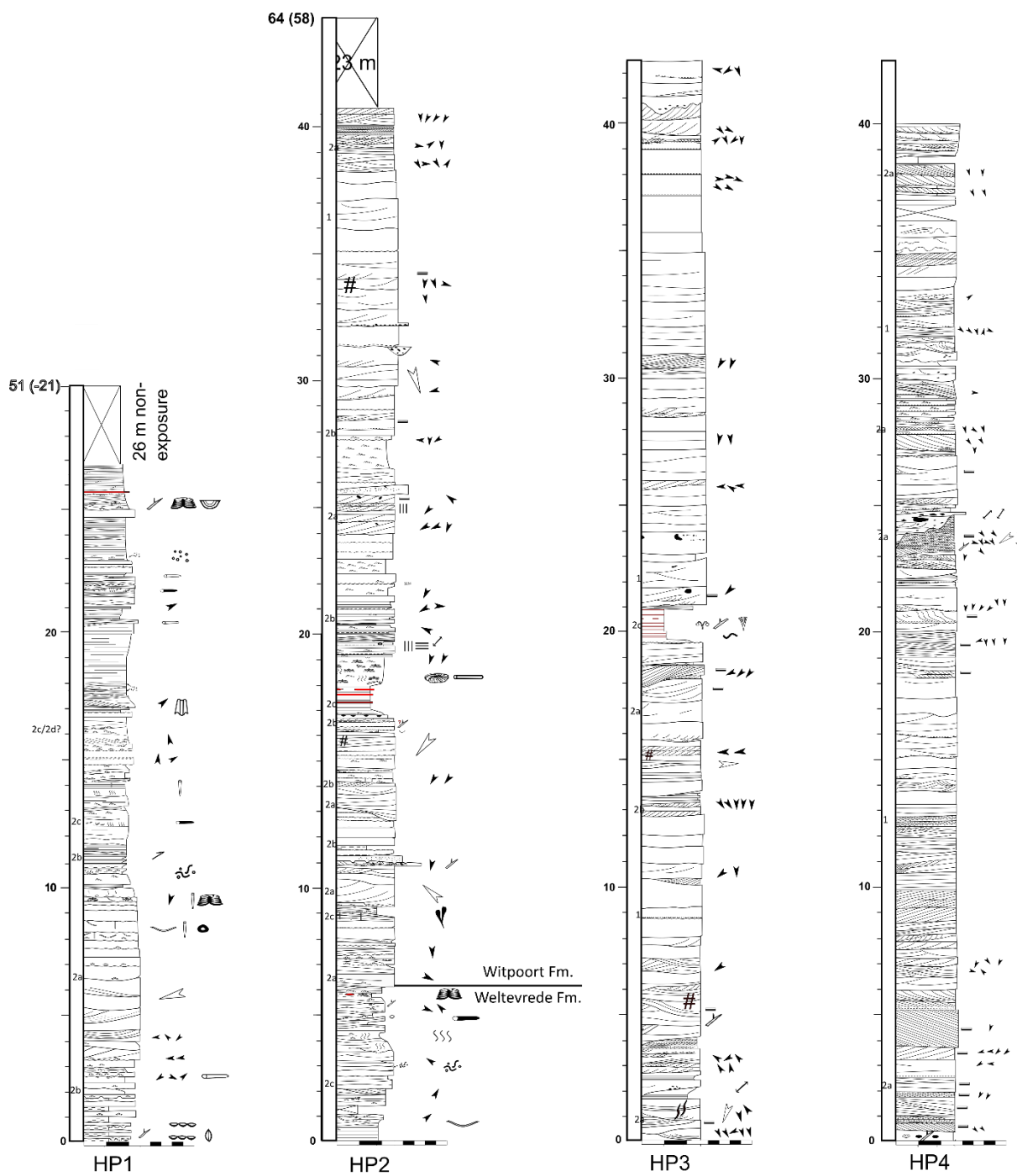
Appendix 1F – Section AD



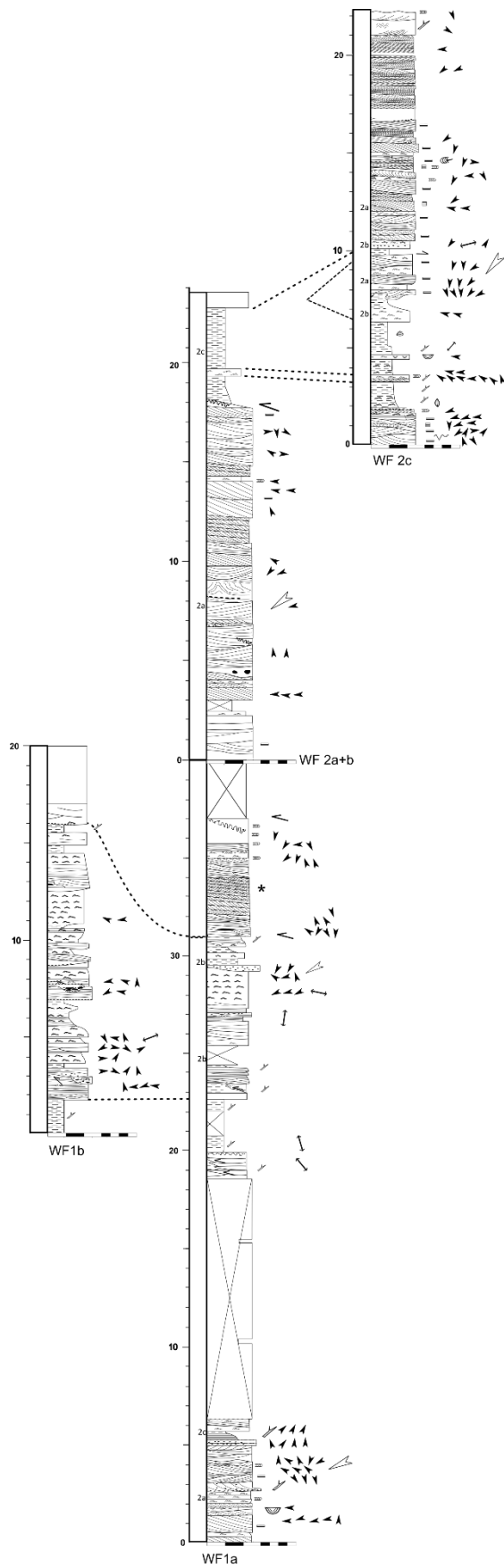
Appendix 1G – Section SF



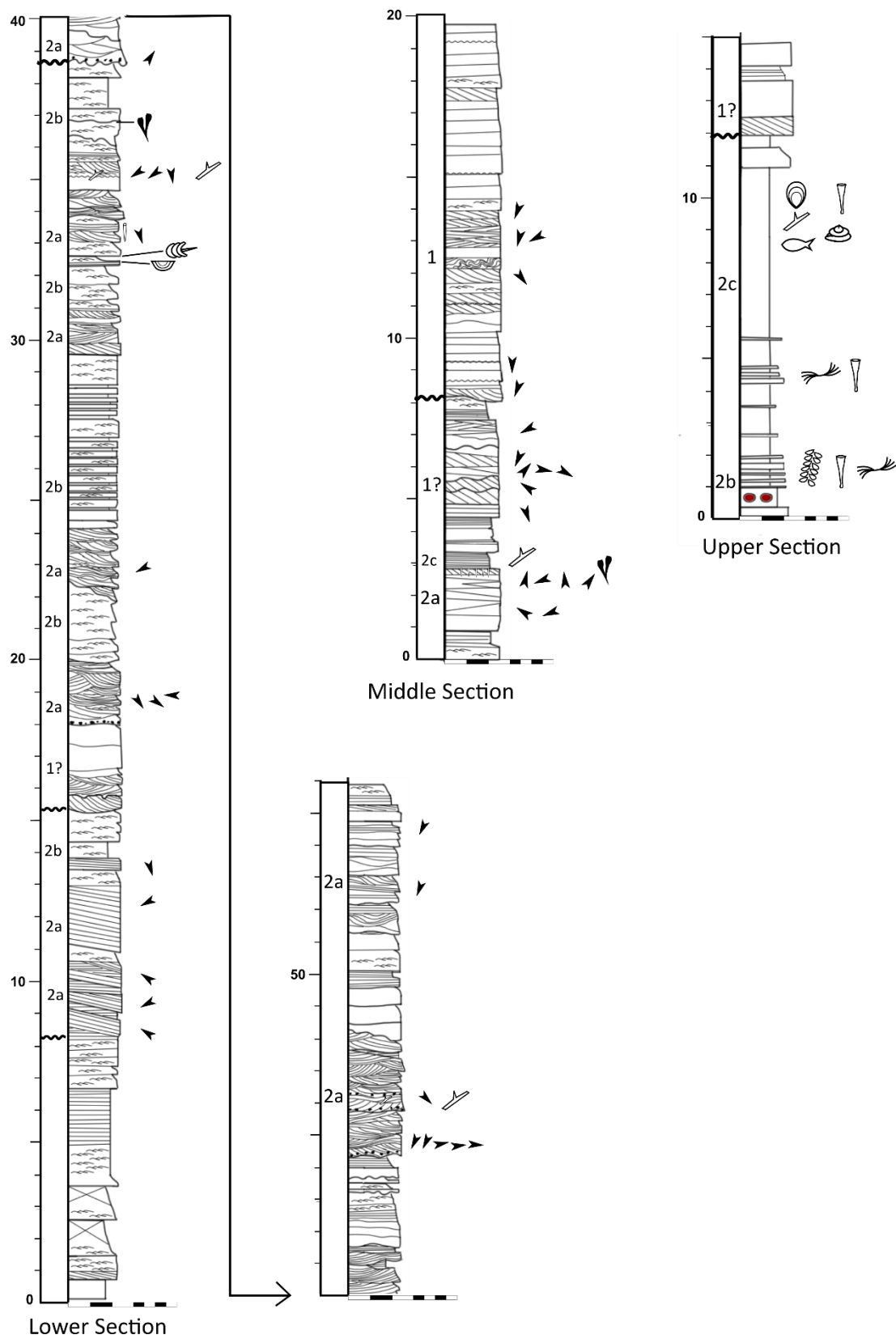
Appendix 1H – Section HP



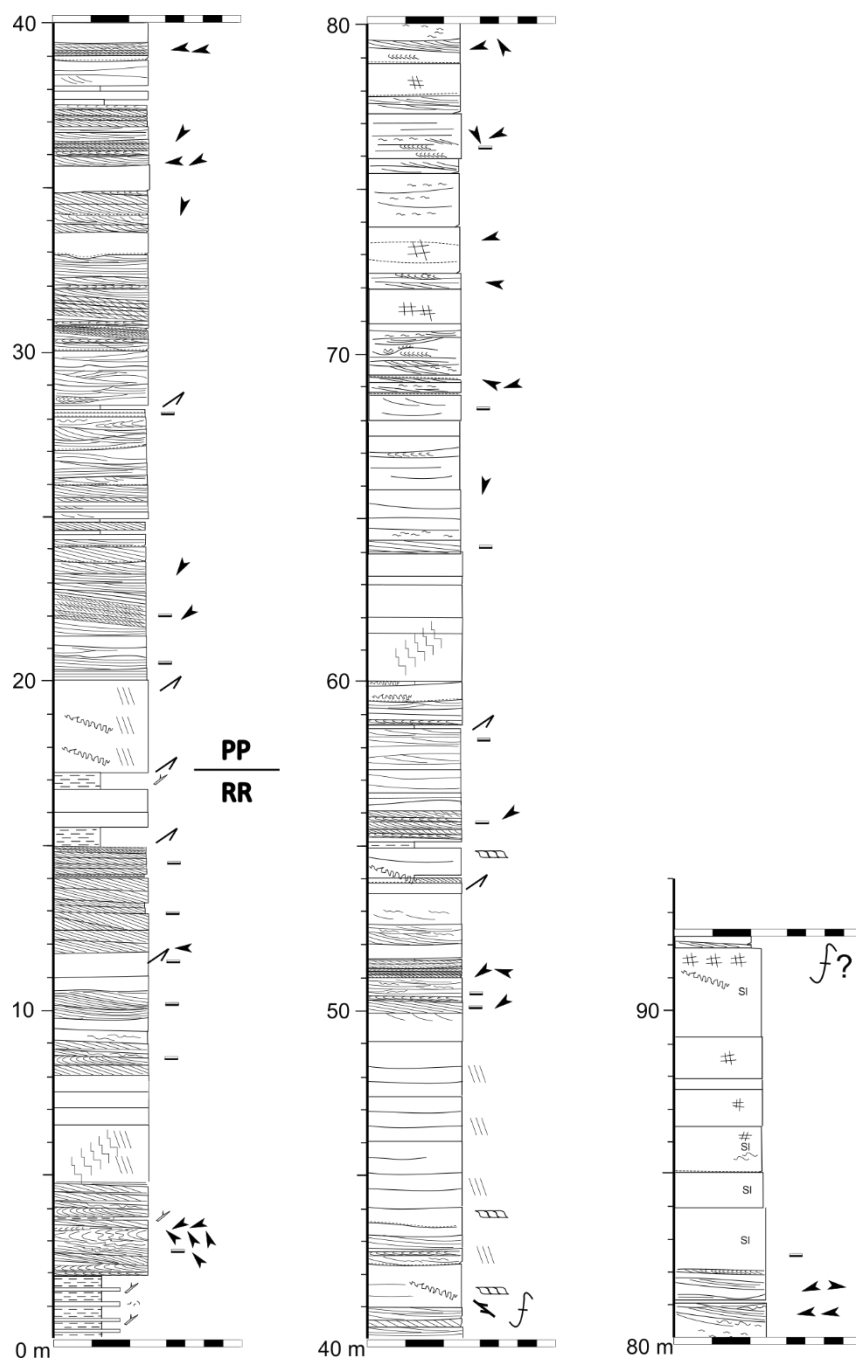
Appendix 1J – Section WF



Appendix 1K – Section RR



Appendix 1L – Section GD



Appendix 2 – Palaeocurrent data

Table 10-1 (next 9 pages): Table of flow vector data (green = upper Weltevrede Formation; yellow = Witpoort Formation; m = metre reference in stratigraphic sections)

section	m	facies	FA	indicator type	bedding		averaged bedding			vector measurement		rotated vector	
					dip°	dip dir.	dip°	dip dir.	plunge	dip°	dip dir.	dip°	dip dir./trend
HP 1	2.5	Sr b	2a	foreset	30	8	35	27	3.7 ESE	45	50	28	113
HP 1		Sr b	2a	foreset	40	25	35	27	3.7 ESE	25	15	12	234
HP 1		Sr b	2a	foreset			35	27	3.7 ESE	50	36	16	48
HP 1	3.5	St p	2a	foreset			35	27	3.7 ESE	30	4	14	265
HP 1		St p	2a	foreset			35	27	3.7 ESE	30	12	10	260
HP 1	4.0	St	2a	foreset	25	10	28	30	3.7 ESE	25	328	27	262
HP 1		St	2a	foreset	30	14	28	28	3.7 ESE	40	358	20	316
HP 1		St	2a	foreset			28	28	3.7 ESE	10	350	22	224
HP 1	5-6.5	St	2a	bed LBS			28	28	3.7 ESE	25	22	5	243
HP 1		St	2a	bed LBS			28	28	3.7 ESE	20	352	17	251
HP 1		St	2a	bed LBS			28	28	3.7 ESE	35	20	8	345
HP 1		St	2a	bed LBS			28	28	3.7 ESE	12	18	17	215
HP 1	9.6	Sr a	2b	foreset	30	38	29	28	3.7 ESE	12	46	18	196
HP 1	15.2	Sp	2d	foreset	30	34	29	28	3.7 ESE	45	20	17	5
HP 1		Sp	2d	foreset			29	28	3.7 ESE	40	40	13	61
HP 1	15.6	St p	2d	foreset			29	28	3.7 ESE	40	52	17	347
HP 1	17.2	St p	2d	foreset	25	32	30	28	3.7 ESE	40	36	12	50
HP 1	21	Sr	2c	ripple foreset	30	35	30	28	3.7 ESE	45	45	19	65
HP 2	1.0	Su	2c	foreset	25	44	20	32	3.7 ESE	40	42	21	49
HP 2	2	Su	2c	foreset			20	32	3.7 ESE	40	32	20	30
HP 2	3	St p	2c	foreset	33	26	33	28	3.7 ESE	35	14	9	304
HP 2	5.5	Su	2c	foreset	37	25	37	27	3.7 ESE	40	50	14	112
HP 2		Su	2c	foreset			37	27	3.7 ESE	45	6	17	316
HP 2	6.2	St p	2b	foreset (apparent dip)			30	28	3.7 ESE	30	18		
HP 2		St p	2b	foreset (apparent dip)			30	28	3.7 ESE	18	90	2	116
HP 2	7.2	St p	2b	foreset (apparent dip)	30	32	30	28	3.7 ESE	16	30		
HP 2		St p	2b	foreset (apparent dip)			30	28	3.7 ESE	11	109	16	176
HP 2	10-11	St	2a	bed LBS			25	30	3.7 ESE	36	14	14	344
HP 2		St	2a	bed LBS			25	30	3.7 ESE	37	20	13	360
HP 2		St	2a	bed LBS			25	30	3.7 ESE	25	27	2	299
HP 2		St	2a	bed LBS			25	30	3.7 ESE	25	0	13	285
HP 2		St	2a	bed LBS			25	30	3.7 ESE	22	357	14	272
HP 2	11.4	Gi	2a	clast imbrication			25	30	3.7 ESE	60	20	36	193
HP 2	13.5	Sp	2a	foreset	25	22	25	30	3.7 ESE	8	2	18	222
HP 2		Sp	2a	foreset			25	30	3.7 ESE	5	295	26	221
HP 2		Sp	2a	foreset			25	30	3.7 ESE	30	15	9	327
HP 2	13.3	St	2a	bed LBS			25	30	3.7 ESE	40	17	17	357
HP 2	13.5	St	2a	bed LBS			25	30	3.7 ESE	38	6	18	333
HP 2	13.8	St	2a	bed LBS			25	30	3.7 ESE	36	22	12	3
HP 2		St	2a	bed LBS			25	30	3.7 ESE	25	356	15	284
HP 2	14.2	St	2a	bed LBS			25	30	3.7 ESE	25	22	4	296
HP 2	14.6	Sp	2a	bed LBS	25	26	25	30	3.7 ESE	33	22	9	357
HP 2	14.3	Sp	2b	foreset	30	29	31	29	3.7 ESE	3	245	29	212
HP 2		Sp	2b	foreset			31	29	3.7 ESE	12	14	20	218
HP 2	15.4	Sp	2b	bed LBS			33	29	3.7 ESE	30	36	4	164
HP 2	15.6	Sp	2b	bed LBS			33	29	3.7 ESE	20	330	28	247
HP 2	16	Sp	2b	bed LBS	35	12	33	29	3.7 ESE	25	324	31	256
HP 2	16.2	Sp	2b	bed LBS			33	29	3.7 ESE	12	24	21	211
HP 2	19	Sr a	2b	foreset			27	29	3.7 ESE	10	50	9	196
HP 2		Sr a	2b	foreset			27	29	3.7 ESE	10	36	17	205
HP 2	20.2	Sr a	2b	foreset			27	29	3.7 ESE	28	8	10	294
HP 2		Sr a	2b	form set crest			27	29	3.7 ESE		SW	0	240
HP 2	21.7	Sr a	2b	foreset			27	29	3.7 ESE	15	32	12	206
HP 2	22	Sr a		rib and furrow			27	29	3.7 ESE	40	30	13	28
HP 2	21.2	St p	2b	foreset			27	29	3.7 ESE	35	52	14	93
HP 2	24	Sp	2a	foreset			27	29	3.7 ESE	14	340	21	240
HP 2		Sp	2a	foreset			27	29	3.7 ESE	17	354	16	246
HP 2		Sp	2a	foreset			27	29	3.7 ESE	8	52	20	199
HP 2	24.9	Sp		parting lineation	28	35	27	29	3.7 ESE	5	113	27	197
HP 2	25	Sp	2a	low angled planar foreset			27	29	3.7 ESE	13	13	15	223
HP 2		Sp	2a	bed LBS			27	29	3.7 ESE	24	24	4	246
HP 2	25.7	Gi		clast tilt			27	29	3.7 ESE	60	43	34	49
HP 2		Gi		clast tilt			27	29	3.7 ESE	40	30	13	28
HP 2	27.8	Sr a	2b	foreset			27	29	3.7 ESE	26	342	21	274
HP 2		Sr a	2b	foreset			24	30	3.7 ESE	10	91	21	184
HP 2		Sr a	2b	foreset			24	30	3.7 ESE	14	7	12	236
HP 2	29.4	Sp	2a	foreset	25	42	24	30	3.7 ESE	19	12	8	255
HP 2	9.6-30.	Sm		bed LBS			24	30	3.7 ESE	20	35	4	189
HP 2		Sm		bed LBS			24	30	3.7 ESE	12	57	14	187
HP 2		Sm		bed LBS			24	30	3.7 ESE	28	54	11	108
HP 2		Sp	2a	foreset			24	30	3.7 ESE	24	6	10	288
HP 2	33.4	St	1	bed LBS	25	37	24	30	3.7 ESE	35	48	14	75
HP 2	33.4	St	1	foreset			24	30	3.7 ESE	14	90	20	173
HP 2	34	St	1	foreset			24	30	3.7 ESE	15	50	12	186
HP 2		St	1	foreset			24	30	3.7 ESE	18	60	12	162
HP 2		St	1	foreset			24	30	3.7 ESE	37	72	24	107
HP 2	38.5	Sl	2a	foreset			21	32	3.7 ESE	20	87	18	150

section	m	facies	FA	indicator type	bedding		averaged bedding			vector measurement		rotated vector	
					dip°	dip dir	dip°	dip dir.	plunge	dip°	dip dir.	dip°	dip dir./trend
HP 2		Sl	2a	foreset			21	32	3.7 ESE	29	55	12	91
HP 2		Sl	2a	foreset			21	32	3.7 ESE	19	68	12	147
HP 2		Sl	2a	foreset			21	32	3.7 ESE	45	35	24	35
HP 2	39.5	St	2a	foreset			21	32	3.7 ESE	40	44	20	53
HP 2		St	2a	foreset			21	32	3.7 ESE	35	68	21	98
HP 2		St	2a	foreset			21	32	3.7 ESE	12	66	13	180
HP 2	40	Sp	2a	foreset	22	42	21	32	3.7 ESE	12	86	17	176
HP 2		Sp	2a	foreset			21	32	3.7 ESE	5	42	16	208
HP 2		Sp	2a	foreset			21	32	3.7 ESE	5	240	26	217
HP 2		Sp	2a	foreset			21	32	3.7 ESE	3	194	22	219
HP 3	0.3	St	2a	foreset	22	46	20	32	3.7 ESE	18	74	14	153
HP 3		St	2a	foreset	22	32	22	32	3.7 ESE	20	80	16	147
HP 3		St	2a	bed LBS	22	40	21	32	3.7 ESE	19	78	15	150
HP 3		St	2a	bed LBS			21	32	3.7 ESE	17	19	6	253
HP 3	0.6	St	2a	foreset			21	32	3.7 ESE	12	354	14	245
HP 3		St	2a	foreset			21	32	3.7 ESE	15	72	13	165
HP 3		St	2a	foreset			21	32	3.7 ESE	16	77	15	161
HP 3	1	St	2a	foreset			21	32	3.7 ESE	15	78	15	165
HP 3	1.6	St	2a	bed LBS			21	32	3.7 ESE	27	12	10	327
HP 3	1.6	St	2a	foreset			21	32	3.7 ESE	28	6	13	321
HP 3	3	Sp	2b	foreset			15	35	3.7 ESE	37	347	29	326
HP 3		Sp	2b	foreset			15	35	3.7 ESE	30	8	18	346
HP 3	3.3	St	2b	foreset			15	35	3.7 ESE	21	349	15	304
HP 3		St	2b	foreset			15	35	3.7 ESE	30	345	23	317
HP 3		St	2b	foreset			15	35	3.7 ESE	20	325	20	282
HP 3	2	St	2b	Mudcast elongation			15	35	3.7 ESE	30-210			228-48
HP 3	3.6	Sp	2b	foreset	15	28	15	35	3.7 ESE	14	300	21	256
HP 3		Sp	2b	foreset	16	36	16	36	3.7 ESE	15	317	19	266
HP 3	4.5	St		foreset			16	36	3.7 ESE	35	26	20	19
HP 3		St		foreset			16	36	3.7 ESE	35	24	21	16
HP 3		St		foreset			16	36	3.7 ESE	18	8	8	312
HP 3	7	St	1	foreset			16	36	3.7 ESE	5	325	14	234
HP 3	7.0	St	1	bed LBS			16	36	3.7 ESE	18	28	4	356
HP 3	10.2	Sp	1	foreset			16	36	3.7 ESE	5	283	18	230
HP 3	10.5	Sl	1	foreset			16	36	3.7 ESE	10	113	16	177
HP 3	12.4	Sm	1	bed LBS			16	36	3.7 ESE	30	54	16	69
HP 3	13	Sp	2b	foreset			16	36	3.7 ESE	16	84	13	144
HP 3		Sp	2b	foreset			16	36	3.7 ESE	10	114	16	177
HP 3		Sp	2b	foreset			16	36	3.7 ESE	17	88	14	143
HP 3	13.2	Sp	2b	foreset			16	36	3.7 ESE	8	123	17	185
HP 3		Sp	2b	foreset			16	36	3.7 ESE	8	96	13	182
HP 3	15	Sl	2a	bed LBS			16	36	3.7 ESE	25	38	10	41
HP 3		Sl	2a	bed LBS			16	36	3.7 ESE	25	22	11	3
HP 3		Sl	2a	bed LBS			16	36	3.7 ESE	30	15	17	357
HP 3		Sl	2a	bed LBS			16	36	3.7 ESE	24	32	9	26
HP 3		Sl	2a	bed LBS			16	36	3.7 ESE	20	28	6	7
HP 3	15.5	St	2a	foreset			16	36	3.7 ESE	15	316	19	266
HP 3		St	2a	foreset			16	36	3.7 ESE	4	348	13	228
HP 3		St	2a	foreset			16	36	3.7 ESE	20	40	5	52
HP 3	18.5	Sp	2a	foreset	23	31	16	36	3.7 ESE	10	298	19	246
HP 3		Sp	2a	foreset			16	36	3.7 ESE	3	304	15	226
HP 3		Sp	2a	foreset			16	36	3.7 ESE	3	245	18	219
HP 3	21.8	St	1	foreset			16	36	3.7 ESE	5	252	19	224
HP 3	25.8	Sp	1	foreset			16	36	3.7 ESE	15	318	19	267
HP 3		Sp	1	foreset			16	36	3.7 ESE	25	335	22	299
HP 3		Sp	1	foreset			16	36	3.7 ESE	16	323	18	272
HP 3	27.5	Sp	1	foreset			16	36	3.7 ESE	12	162	24	190
HP 3		Sp	1	foreset			16	36	3.7 ESE	12	122	19	174
HP 3	30.6	Sp	1	foreset	15	30	15	35	3.7 ESE	10	38	5	210
HP 3		Sp	1	foreset			15	35	3.7 ESE	4	89	13	200
HP 3	37.0	Sp	2a	foreset			7	50	3.7 ESE	24	109	21	124
HP 3		Sp	2a	foreset			7	50	3.7 ESE	20	100	16	118
HP 3	37.8	Sp	2a	foreset	15	85	7	50	3.7 ESE	25	82	19	92
HP 3		Sp	2a	foreset			7	50	3.7 ESE	24	89	19	101
HP 3		Sp	2a	foreset			7	50	3.7 ESE	28	101	24	113
HP 3	39.4	Sp	2a	parabolic foreset			7	50	3.7 ESE	15	52	8	53
HP 3		Sp	2a	parabolic foreset			7	50	3.7 ESE	15	131	16	156
HP 3		Sp	2a	parabolic foreset			7	50	3.7 ESE	12	232	19	231
HP 3		Sp	2a	parabolic foreset			7	50	3.7 ESE	16	165	20	183
HP 3	40.0	St	2a	foreset	15	50	7	50	3.7 ESE	19	116	17	136
HP 3		St	2a	foreset			7	50	3.7 ESE	19	98	15	117
HP 3	41.2	Sp	2a	bed LBS			7	50	3.7 ESE	5	184	11	211
HP 3	42.2	St	2a	foreset	5	80	7	50	3.7 ESE	16	305	19	284
HP 3		St	2a	foreset			7	50	3.7 ESE	4	242	11	234
HP 3		St	2a	foreset			7	50	3.7 ESE	15	135	16	160
HP 4	0.4	Sp	2a	foreset	8	45	8	44	3.7 ESE	15	100	12	131
HP 4		Sp	2a	foreset			8	44	3.7 ESE	16	106	14	135
HP 4	1.9	Sp	2a	foreset			8	44	3.7 ESE	10	192	17	206
HP 4		Sp	2a	foreset			8	44	3.7 ESE	5	194	13	212
HP 4	3	Sl p	2a	foreset			8	44	3.7 ESE	5	256	13	236

section	m	facies	FA	indicator type	bedding		averaged bedding			vector measurement		rotated vector	
					dip°	dip dir	dip°	dip dir.	plunge	dip°	dip dir.	dip°	dip dir./trend
HP 4		Sl p	2a	foreset			8	44	3.7 ESE	10	305	14	270
HP 4	3.6	Sp	2a	foreset			8	44	3.7 ESE	10	279	16	255
HP 4		Sp	2a	foreset			8	44	3.7 ESE	17	284	22	266
HP 4		Sl p	2a	foreset			8	44	3.7 ESE	5	225	13	224
HP 4		Sl p	2a	foreset			8	44	3.7 ESE	5	236	13	228
HP 4	5	Sp	2a	foreset	5	28	5	55	3.7 ESE	8	174	11	196
HP 4	6.8	St	1	foreset	14	41	8	44	3.7 ESE	16	32	8	20
HP 4		St	1	foreset			8	44	3.7 ESE	12	358	9	316
HP 4		St	1	foreset			8	44	3.7 ESE	15	78	9	105
HP 4	7	St	1	foreset			8	44	3.7 ESE	10	190	17	204
HP 4	7.2	St	1	foreset			8	44	3.7 ESE	17	112	16	139
HP 4		St	1	foreset	13	68	9	44	3.7 ESE	3	114	8	202
HP 4	19.6	Sl p	1	foreset	5	22	5	54	3.7 ESE	2	288	6	249
HP 4		Sl p	1	foreset			5	54	3.7 ESE	5	115	5	174
HP 4		Sl p	1	foreset			5	54	3.7 ESE	5	106	4	170
HP 4		Sl p	1	foreset			5	54	3.7 ESE	4	100	4	181
HP 4	22.7	SSDS	1	Axis of steepening	5	50	4	59	3.7 ESE			0	191
HP 4	23.5	Sp	2a	foreset			4	59	3.7 ESE	18	126	17	138
HP 4		Sp	2a	toeset			4	59	3.7 ESE	11	112	9	132
HP 4	23	Sp	2a	foreset			4	59	3.7 ESE	10	112	8	134
HP 4	21	St	1	foreset			4	59	3.7 ESE	15	170	17	182
HP 4		St	1	foreset			4	59	3.7 ESE	10	178	12	194
HP 4		St	1	foreset			4	59	3.7 ESE	8	204	12	215
HP 4		St	1	foreset			4	59	3.7 ESE	14	223	18	226
HP 4	21.2	St	1	foreset			4	59	3.7 ESE	15	152	16	166
HP 4		St	1	foreset			4	59	3.7 ESE	17	168	19	179
HP 4	23.5	St	2a	Trough axis			4	59	3.7 ESE	8-188			8-188
HP 4		St	2a	bed LBS			4	59	3.7 ESE	17	106	15	117
HP 4		St	2a	bed LBS			4	59	3.7 ESE	24	342	23	333
HP 4	24.8	St	1	gutter cast			4	59	3.7 ESE	46-226			45-225
HP 4	23.3	Sp	2a	bed lbs			4	59	3.7 ESE	17	95	14	104
HP 4		Sp	2a	foreset			4	59	3.7 ESE	21	145	21	155
HP 4	23.5	Sp	2a	bed LBS			4	59	3.7 ESE	17	100	14	110
HP 4		Sp	2a	foreset			4	59	3.7 ESE	25	138	25	146
HP 4	27	Gi	1	Mudclast elongation			4	59	3.7 ESE	34-214			34-214
HP 4	27.5	Sp	2a	foreset			4	59	3.7 ESE	15	129	14	144
HP 4		Sp	2a	foreset			4	59	3.7 ESE	17	135	17	148
HP 4		Sp	2a	foreset			4	59	3.7 ESE	19	168	21	178
HP 4	28	St	2a	foreset			4	59	3.7 ESE	20	156	21	166
HP 4		St	2a	foreset			4	59	3.7 ESE	13	110	11	126
HP 4		St	2a	foreset			4	59	3.7 ESE	19	152	20	163
HP 4	29.6	Sp	2b	foreset	7	244	4	59	3.7 ESE	16	86	13	94
HP 4	33.3	SSDS	1	Parabolic foreset			4	59	3.7 ESE	28	54	24	53
HP 4	32	Sl	1	foreset			4	59	3.7 ESE	14	115	12	130
HP 4		Sl	1	foreset			4	59	3.7 ESE	22	123	21	132
HP 4		Sl	1	foreset			4	59	3.7 ESE	18	152	19	164
HP 4		Sl	1	foreset			4	59	3.7 ESE	21	19	18	11
HP 4		Sl	1	foreset			4	59	3.7 ESE	16	102	13	113
HP 4	37.2	St	2a	foreset			4	59	3.7 ESE	11	136	11	157
HP 4		St	2a	foreset			4	59	3.7 ESE	13	145	13	162
HP 4	37.2	St	2a	bed LBS	3	7	4	59	3.7 ESE	10	280	13	269
HP 4	38.2	Sp	2a	bed LBS			4	59	3.7 ESE	10	235	14	236
HP 4		Sp	2a	bed LBS			4	59	3.7 ESE	13	206	17	213
HP 4	38.2	Sp	2a	foreset			4	59	3.7 ESE	10	142	10	164
HP 4		Sp	2a	foreset			4	59	3.7 ESE	8	152	9	178
WF 1a	0.0	Sp	2a	foreset			11	1	3.7 WSW	22	296	20	267
WF 1a	1.0	St	2a	foreset	10	19	11	1	3.7 WSW	20	284	20.5	253
WF 1a		St	2a	foreset			11	1	3.7 WSW	17	298	15.5	259
WF 1a		St	2a	foreset			11	1	3.7 WSW	18	295	16.8	259
WF 1a		St	2a	foreset			11	1	3.7 WSW	36	2	25	2
WF 1a		St	2a	ridges and troughs			11	1	3.7 WSW			0	273.1
WF 1a	1.5	St	2a	rib and furrow			11	1	3.7 WSW			0	271.3
WF 1a	3.5	St	2a	foreset			11	1	3.7 WSW	25	321	18	298
WF 1a		St	2a	foreset			11	1	3.7 WSW	21	326	13.6	298
WF 1a		St	2a	foreset			11	1	3.7 WSW	25	330	16.6	310
WF 1a	3.2	St	2a	foreset			11	1	3.7 WSW	9	94	14.5	142
WF 1a		St	2a	foreset			11	1	3.7 WSW	6	135	15.7	164
WF 1a		St	2a	foreset			11	1	3.7 WSW	3	72	10.4	165
WF 1a	4.0	St	2a	foreset			11	1	3.7 WSW	24	0	13	358
WF 1a		St	2a	foreset			11	1	3.7 WSW	25	322	17.8	300
WF 1a		St	2a	foreset			11	1	3.7 WSW	32	344	21.7	335
WF 1a	4.2	St	2a	rib and furrow			11	1	3.7 WSW	200		0	20.3
WF 1a		St	2a	rib and furrow			11	1	3.7 WSW	233		0	53.3
WF 1a	4.5	St	2a	current parallel ridge			11	1	3.7 WSW	66-246		0	66.3
WF 1a	5.0	Gi	2c	Clast inclination	11	20	11	1	3.7 WSW	340		0	340.3
WF 1a		Gi	2c	Clast inclination			11	1	3.7 WSW	40		0	40.3
WF 1a		Gi	2c	imbrication			11	1	3.7 WSW	5		0	5.3
WF 1a		Gi	2c	scour tails			11	1	3.7 WSW	0		0	0.3
WF 1a	5.7	St	2d	foreset			11	1	3.7 WSW	30	25	20.4	46

section	m	facies	FA	indicator type	bedding		averaged bedding			vector measurement		rotated vector	
					dip°	dip dir	dip°	dip dir.	plunge	dip°	dip dir.	dip°	dip dir./trend
WF 1a		St	2d	foreset			11	1	3.7 WSW	30	19	19.8	27
WF 1a	5.9	Sl p	2d	foreset			11	1	3.7 WSW	20	14	9.6	27
WF 1a	20.4	Fh	2c	plant alignment			3	257	3.7 WSW	140-320		0	319.8
WF 1a	19.5	Fh	2c	plant alignment			3	257	3.7 WSW	164-344		0	343.8
WF 1a	27.0	Sl	2b	Gutter	3	168	3	257	3.7 WSW	188-8		0	187.8
WF 1a	28.0	Sr	2b	foreset			3	257	3.7 WSW	15	246	12.1	243
WF 1a		Sr	2b	foreset			3	257	3.7 WSW	21	256	18	256
WF 1a		Sr	2b	foreset			3	257	3.7 WSW	25	238	22.2	235
WF 1a		Sr	2b	form set crest			3	257	3.7 WSW	192-12		0	282
WF 1a	29.0	Sr	2b	foreset			4	321	3.7 WSW	18	309	14.1	305
WF 1a		Sr	2b	foreset			4	321	3.7 WSW	13	271	10.9	255
WF 1a		Sr	2b	foreset			4	321	3.7 WSW	20	337	16.2	341
WF 1a	29.2	St	2b	LBS	3	342	4	321	3.7 WSW	20	156	23.9	153
WF 1a		St	2b	LBS			4	321	3.7 WSW	15	326	11	328
WF 1a		Sl	2b	foreset			4	321	3.7 WSW	17	220	18.2	208
WF 1a		Sl	2b	foreset			4	321	3.7 WSW	20	220	21.1	210
WF 1a	31.8	Su	2a	foreset	10	302	9	355	3.7 WSW	12	340	4.1	304
WF 1a		Su	2a	foreset	10	330	9	355	3.7 WSW	12	236	18.1	210
WF 1a		Su	2a	foreset			9	355	3.7 WSW	4	40	6.7	150
WF 1a		Su	2a	foreset			9	355	3.7 WSW	3	336	6.2	184
WF 1a	31.5	Sl p	2a	foreset			9	355	3.7 WSW	20	346	11.2	338
WF 1a		Sl p	2a	foreset			9	355	3.7 WSW	26	334	17.9	324
WF 1a	32.0	St	2a	foreset			9	355	3.7 WSW	6	140	14.3	161
WF 1a	32.5	Sl p	2a	LBS			9	355	3.7 WSW	18	0	9.1	364
WF 1a	35.0	St	2a	rib and furrow			9	355	3.7 WSW	230		0	49.4
WF 1a		St	2a	rib and furrow			9	355	3.7 WSW	200		0	19.4
WF 1a	34.5	Sp	2a	LBS			9	355	3.7 WSW	25	336	16.8	326
WF 1a		Sp	2a	foreset			9	355	3.7 WSW	20	350	11.1	345
WF 1a	35.0	St	2a	foreset	10	10	9	355	3.7 WSW	21	346	12.2	339
WF 1a		St	2a	foreset			9	355	3.7 WSW	22	346	13.2	339
WF 1a	35.5	St	2a	rib and furrow			9	355	3.7 WSW	216		0	35.4
WF 1a		St	2a	foreset			9	355	3.7 WSW	18	295	15.6	265
WF 1a		St	2a	foreset			9	355	3.7 WSW	5	113	12.1	153
WF 1a	36.0	St	2a	Rib and furrow			9	355	3.7 WSW	190		0	9.4
WF 1b	2.4	Sr	2b	foreset	8	352	13	2	3.7 WSW	24	355	11.3	346
WF 1b		Sr	2b	foreset	14	20	13	2	3.7 WSW	15	324	9.3	265
WF 1b		Sr	2b	foreset			13	2	3.7 WSW	20	288	20.6	251
WF 1b		Sr	2b	foreset			13	2	3.7 WSW	16	332	8.1	278
WF 1b	3.4	Sl	2b	foreset			13	2	3.7 WSW	27	60	22.6	86
WF 1b		Sl	2b	foreset			13	2	3.7 WSW	14	60	12.9	116
WF 1b		Sl	2b	foreset			13	2	3.7 WSW	16	10	3.6	37
WF 1b		Sl	2b	foreset			13	2	3.7 WSW	20	356	7.2	344
WF 1b	3.8	Sr	2b	foreset			13	2	3.7 WSW	15	28	6.4	86
WF 1b		Sr	2b	foreset			13	2	3.7 WSW	25	18	13	32
WF 1b	4.5	Sl	2b	foreset			13	2	3.7 WSW	15	14	3.4	61
WF 1b		Sl	2b	foreset			13	2	3.7 WSW	11	52	10.2	126
WF 1b		Sl	2b	foreset			13	2	3.7 WSW	15	42	9.6	99
WF 1b		Sl	2b	foreset			13	2	3.7 WSW	15	286	17.3	240
WF 1b		Sr b	2b	Crestline			13	2	3.7 WSW	340-160		0	69
WF 1b		Sr b	2b	foreset			13	2	3.7 WSW	30	350	17.5	341
WF 1b		Sr b	2b	foreset			13	2	3.7 WSW	20	320	13.5	280
WF 1b		Sr b	2b	foreset			13	2	3.7 WSW	8	30	6.9	149
WF 1b	7.3	Sp	2b	foreset	15	355	13	2	3.7 WSW	12	304	12.1	239
WF 1b	7.8	Sr	2b	foreset			13	2	3.7 WSW	24	312	18.5	280
WF 1b	8.0	Sr	2b	foreset			13	2	3.7 WSW	15	320	10.2	261
WF 1b		Sr	2b	foreset			13	2	3.7 WSW	20	328	11.8	290
WF 1b		Sr	2b	foreset			13	2	3.7 WSW	30	0	17	357
WF 1b	11.0	Sr	2b	foreset			13	2	3.7 WSW	30	316	22.9	293
WF 1b		Sr	2b	foreset			13	2	3.7 WSW	23	296	21.2	263
WF 1b	16.0	Sl	2b	LBS			13	2	3.7 WSW	30	3	17	3
WF 1b		Sl	2b	LBS			13	2	3.7 WSW	25	24	13.7	43
WF 2a	3.4	Sp	2a	foreset			8	354	3.7 WSW	25	285	23.3	267
WF 2a		Sp	2a	foreset			8	354	3.7 WSW	25	284	23.5	266
WF 2a		Sp	2a	foreset			8	354	3.7 WSW	25	294	22.1	276
WF 2a	5.4	St	2a	Lateral accretion set?	7	20	8	354	3.7 WSW	22	354	17	353
WF 2a		St	2a	Lateral accretion set?			8	354	3.7 WSW	26	0	18.1	2
WF 2a	8.0	St	2a	Current parr ridges			10	0	3.7 WSW	55-235		0	54.4
WF 2a	7.8	St	2a	foreset			10	0	3.7 WSW	16	294	15	257
WF 2a	9.5	Sl	2a	foreset			10	0	3.7 WSW	10	260	15.3	220
WF 2a		Sl	2a	foreset			10	0	3.7 WSW	30	262	32.8	246
WF 2a	10.0	Sp	2a	foreset			10	0	3.7 WSW	25	318	18.8	298
WF 2a	12.4	St	2a	LBS			10	0	3.7 WSW	20	8	10.2	15
WF 2a		St	2a	foreset			10	0	3.7 WSW	20	350	10.3	338
WF 2b	13.5	Sp	2a	LBS			10	0	3.7 WSW	33	356	23.1	354
WF 2b		Sp	2a	foreset			10	0	3.7 WSW	20	310	15.6	281
WF 2b		Sp	2a	foreset			10	0	3.7 WSW	25	294	22.8	272
WF 2b	14.0	St	2a	rib and furrow	12	4	12	4	3.7 WSW	264		0	273
WF 2b	15.5	St	2a	LBS			12	4	3.7 WSW	25	326	17.2	301
WF 2b		St	2a	LBS			12	4	3.7 WSW	15	42	9.1	94
WF 2b	16.8	St	2a	foreset			20	212	3.7 WSW	22	128	27.9	84

section	m	facies	FA	indicator type	bedding		averaged bedding			vector measurement		rotated vector	
					dip°	dip dir	dip°	dip dir.	plunge	dip°	dip dir.	dip°	dip dir./trend
WF 2b	16.4	St	2a	foreset			20	212	3.7 WSW	22	188	9	123
WF 2b		St	2a	foreset	20	212	20	212	3.7 WSW	25	204	6	174
WF 2c	0.2	St	2a	foreset			3	284	3.7 WSW	28	338	26.3	343
WF 2c		St	2a	foreset			3	284	3.7 WSW	10	156	12.1	145
WF 2c	0.5	St	2a	foreset			3	284	3.7 WSW	10	246	7.9	232
WF 2c		St	2a	foreset			3	284	3.7 WSW	22	22	22.6	29
WF 2c	0.7	St	2a	foreset			3	284	3.7 WSW	20	80	22.8	83
WF 2c		St	2a	foreset			3	284	3.7 WSW	15	35	16.3	44
WF 2c	1.0	St	2a	foreset	5	284	3	284	3.7 WSW	12	194	12.4	180
WF 2c	1.0	St	2a	Parabolic foreset			3	284	3.7 WSW	24	254	21.5	250
WF 2c	1.2	St	2a	Parabolic foreset			6	259	3.7 WSW	24	268	29.9	267
WF 2c	1.5	St	2a	foreset			6	259	3.7 WSW	28	248	33.9	250
WF 2c		St	2a	foreset			6	259	3.7 WSW	16	251	21.9	253
WF 2c		St	2a	foreset	10	232	6	259	3.7 WSW	25	264	31	264
WF 2c	1.6	St	2a	rib and furrow			6	259	3.7 WSW	252		0	251.6
WF 2c	3.5	St	2a	rib and furrow			6	259	3.7 WSW	302		0	301.6
WF 2c		St	2a	rib and furrow			6	259	3.7 WSW	304		0	303.6
WF 2c	3.2	St	2a	foreset			6	259	3.7 WSW	30	296	34.9	291
WF 2c		St	2a	foreset			6	259	3.7 WSW	32	270	37.9	269
WF 2c		St	2a	foreset			6	259	3.7 WSW	34	308	38.2	302
WF 2c		St	2a	foreset			6	259	3.7 WSW	32	318	35.4	311
WF 2c		St	2a	foreset			6	259	3.7 WSW	30	350	30.4	340
WF 2c	3.3	Sr	2a	foreset			6	259	3.7 WSW	20	290	25.3	283
WF 2c		Sr	2a	foreset			6	259	3.7 WSW	26	270	31.9	269
WF 2c	4.5	Sr	2a	foreset	10	272	6	259	3.7 WSW	25	280	30.7	277
WF 2c	4.8	Fh	2c	Plant alignment			6	259	3.7 WSW	222-142		0	221.6
WF 2c	6.7	Sr	2c	foreset			6	259	3.7 WSW	24	290	29.3	285
WF 2c		Sr	2c	foreset			6	259	3.7 WSW	25	292	30.2	287
WF 2c	7.8	Sl	2c	foreset			6	259	3.7 WSW	18	142	16.2	161
WF 2c		Sl	2c	Parabolic foreset			6	259	3.7 WSW	12	122	8.7	150
WF 2c		St	2c	foreset			6	259	3.7 WSW	25	165	25.3	178
WF 2c		St	2c	foreset			6	259	3.7 WSW	20	210	24.3	221
WF 2c		St	2c	foreset			6	259	3.7 WSW	20	234	25.6	240
WF 2c	8.2	Sr a	2b	foreset			6	259	3.7 WSW	16	164	16.6	185
WF 2c		Sr a	2b	foreset			6	259	3.7 WSW	6	114	3.6	187
WF 2c	8.7	Sl	2a	foreset			6	259	3.7 WSW	25	200	28.5	210
WF 2c		Sl	2a	foreset			6	259	3.7 WSW	25	196	28.2	207
WF 2c		Sl	2a	foreset			6	259	3.7 WSW	26	216	30.6	223
WF 2c	9.0	St	2a	LBS			6	259	3.7 WSW	16	160	16.2	181
WF 2c		St	2a	LBS			6	259	3.7 WSW	26	232	31.5	237
WF 2c		Sr	2a	foreset			6	259	3.7 WSW	15	174	16.6	195
WF 2c	9.5	Sl	2a	foreset			6	259	3.7 WSW	26	200	29.5	210
WF 2c	10.2	Sr b	2b	Crest	6	218	5	234	3.7 WSW		166-346	0	
WF 2c		Sr b	2b	foreset			5	234	3.7 WSW	20	222	15.2	218
WF 2c		Sr b	2b	foreset			5	234	3.7 WSW	15	25	19.5	32
WF 2c	12.0	Sp	2a	foreset	6	246	5	234	3.7 WSW	25	272	21.3	280
WF 2c		Sp	2a	foreset			5	234	3.7 WSW	25	264	20.8	270
WF 2c	13.0	Sp	2a	foreset			5	234	3.7 WSW	26	232	21	231
WF 2c	13.9	St	2a	Rib and furrow			5	234	3.7 WSW			0	204.7
WF 2c	14.0	St	2a	Oversteepened forset			5	234	3.7 WSW	25	260	20.6	265
WF 2c	13.8	Sp	2a	foreset			5	234	3.7 WSW	25	160	24.1	149
WF 2c	14.3	Sp	2a	foreset			5	234	3.7 WSW	20	206	15.8	197
WF 2c		Sp	2a	foreset			5	234	3.7 WSW	10	116	13.1	96
WF 2c	15.0	Sp	2a	foreset			5	234	3.7 WSW	28	202	23.9	196
WF 2c	15.8	Sp	2a	foreset			5	234	3.7 WSW	26	232	21	231
WF 2c	22.3	St	2a	Large rib and furrow			5	234	3.7 WSW			0	159.7
WF 2c	19.5	Sp	2a	foreset			5	234	3.7 WSW	34	230	29	229
WF 2c		Sp	2a	foreset	3	262	5	234	3.7 WSW	35	245	30.1	246
WF 2c	20.5	Sp	2a	foreset	8	212	5	234	3.7 WSW	20	260	15.7	267
WF 2c	21.2	St	2a	foreset			5	234	3.7 WSW	14	142	15	123
WF 2c	25.0	Sp	2a	foreset			5	234	3.7 WSW	20	134	21.4	121
WF 2c		Sp	2a	foreset			5	234	3.7 WSW	30	158	29.2	149
WF 2c	30.6	Sp	2a	LBS			5	234	3.7 WSW	22	214	17.4	208
WF 2c	32.6	St	2a	foreset			5	234	3.7 WSW	40	180	37.3	174
GD	2.4	Sp	1	foreset	12	215	12	215	3.7 WSW	12	246	6.3	319
GD	3.0	Sp	1	Parabolic foreset	10	250	12	215	3.7 WSW	24	286	23	313
GD		Sp	1	Parabolic axis of def			12	215	3.7 WSW	10		0	328
GD		Sp	1	Parabolic foreset			12	215	3.7 WSW	25	318	30	339
GD	3.5	Sp	1	foreset			12	215	3.7 WSW	40	235	29	241
GD		Sp	1	foreset			12	215	3.7 WSW	35	240	25	250
GD	12.0	Sp	1	foreset	40	204	40	203	3.7 WSW	55	235	27	269
GD	23.5	Sl p	1	foreset	16	215	16	211	3.7 WSW	22	212	6	212
GD	22.5	Sp	1	foreset	23	196	16	211	3.7 WSW	26	220	11	231
GD	34.6	Sp	1	foreset	14	192	17	209	3.7 WSW	30	204	13	196
GD	35.8	Sp	1	foreset	20	190	17	209	3.7 WSW	30	238	17	263
GD		Sp	1	foreset			17	209	3.7 WSW	35	232	20	248
GD	36.5	Sp	1	Parabolic foreset			17	209	3.7 WSW	35	214	18	217
GD	39.3	St	1	foreset			17	209	3.7 WSW	30	236	17	260
GD		St	1	foreset			17	209	3.7 WSW	30	236	17	260
GD	50.2	Sl p	1	foreset	35	205	35	204	3.7 WSW	46	214	12	234

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section	m	facies	FA	indicator type	bedding		averaged bedding			vector measurement		rotated vector	
					dip°	dip dir	dip°	dip dir.	plunge	dip°	dip dir.	dip°	dip dir./trend
GD	51.0	Sp	1	foreset	35	200	35	204	3.7 WSW	40	230	16	285
GD		Sp	1	foreset			35	204	3.7 WSW	40	210	6	235
GD	56.0	Sp	1	foreset	25	208	25	206	3.7 WSW	38	216	14	234
GD	66.0	Sl	1	foreset	34	212	30	205	3.7 WSW	36	204	6	193
GD	69.0	Sp	1	foreset			30	205	3.7 WSW	32	226	11	292
GD		Sp	1	foreset			30	205	3.7 WSW	38	220	11	254
GD	72.2	Sp	1	foreset	30	194	30	205	3.7 WSW	35	226	12	278
GD	73.5	Sl	1	foreset			30	205	3.7 WSW	35	216	7	255
GD	76.5	St	1	foreset			30	205	3.7 WSW	35	196	7	153
GD		St	1	foreset			30	205	3.7 WSW	36	215	8	247
GD	79.5	Sl p	1	foreset			30	205	3.7 WSW	35	216	7	255
GD		Sl p	1	foreset			30	205	3.7 WSW	25	236	15	327
GD	80.8	Sl p	1	foreset	32	184	30	205	3.7 WSW	35	218	8	261
GD		Sl p	1	foreset			30	205	3.7 WSW	36	220	10	262
GD	81.4	Sl	1	foreset			30	205	3.7 WSW	32	210	3	251
GD		Sl	1	foreset			30	205	3.7 WSW	30	167	19	97
WP 1	7.0	Sr b	2c	foreset	60	10	65	15	12 WNW	80	188	30	29
WP 1		Sr b	2c	foreset	70	20	65	15	12 WNW	80	184	30	21
WP 1		Sr b	2c	foreset			65	15	12 WNW	60	28	24	142
WP 1		Sr b	2c	foreset			65	15	12 WNW	85	10	16	42
WP 1		Sr b	2c	foreset			65	15	12 WNW	75	10	7	70
WP 1		Sr b	2c	foreset			65	15	12 WNW	56	4	14	203
WP 1	19.5	Sl	2a	foreset	70	24	70	24	12 WNW	55	22	22	160
WP 1		St	2a	foreset			70	24	12 WNW	62	28	23	137
WP 1	26.5	Sl	2a	foreset	75	18	75	15	12 WNW	65	0	7	239
WP 1		Sr	2a	foreset			75	15	12 WNW	74	0	6	333
WP 1	27.5	Sl	2a	foreset			75	15	12 WNW	80	354	14	334
WP 1		Sl	2a	foreset			75	15	12 WNW	85	0	15	5
WP 1	33.0	Sl	2a	foreset			75	15	12 WNW	80	20	18	81
WP 1		Sl	2a	Topset			75	15	12 WNW	65	16	12	140
WP 1	35.0	Sc	2a	Reactivation			75	15	12 WNW	85	180	25	11
WP 1		Sc	2a	foreset			75	15	12 WNW	66	22	17	129
WP 1	37.5	Sc	2d	IHS forset			75	15	12 WNW	80	2	10	8
WP 1		Sc	2d	IHS forset			75	15	12 WNW	65	10	7	158
WP 1	38.7	Sc	2d	IHS forset			75	15	12 WNW	75	15	11	88
WP 1		Sc	2d	IHS forset			75	15	12 WNW	82	14	15	61
WP 1		Sc	2d	IHS forset			75	15	12 WNW	80	4	10	19
WP 1		Sc	2d	IHS forset			75	15	12 WNW	65	2	6	224
WP 1		Sc	2d	IHS forset			75	15	12 WNW	80	2	10	8
WP 1	44.8	Su	2c	Ripple forset			75	15	12 WNW	60	2	10	213
WP 1		Su	2c	Ripple forset			75	15	12 WNW	70	4	0	265
WP 1		Sr a	2c	Ripple forset	80	15	75	15	12 WNW	85	10	16	42
WP 1	46.0	Sl p	2a	foreset			75	15	12 WNW	65	356	9	256
WP 1	67.0	St	2d	foreset			75	15	12 WNW	85	190	25	35
WP 1	69.0	St	2d	foreset			75	15	12 WNW	88	358	19	1
WP 1		St	2d	foreset			75	15	12 WNW	80	177	31	7
WP 1	71.0	Sl	2b	foreset			75	15	12 WNW	80	16	15	72
WP 1	73.0	Sl	1	foreset	80	8	75	15	12 WNW	65	16	12	140
WP 1		Sl	1	foreset			75	15	12 WNW	60	356	13	237
WP 1		Sl	1	foreset			75	15	12 WNW	75	358	8	329
WP 1		Sl	1	Topset			75	15	12 WNW	70	4	0	265
WP 1	80.0	Sl	2a	foreset			75	15	12 WNW	63	2	8	218
WP 1	81.0	Sl	2d	foreset			75	15	12 WNW	77	358	9	339
WP 1		Sl	2d	foreset			75	15	12 WNW	60	0	11	222
WP 1	85.5	St	2a	foreset	70	16	75	15	12 WNW	50	0	21	211
WP 1		St	2a	foreset			75	15	12 WNW	49	14	23	183
WP 1		St	2a	foreset			75	15	12 WNW	53	22	23	164
WP 1	86.5	Sp	2b	foreset			75	15	12 WNW	60	15	14	160
WP 1	88.0	Sr	2b	foreset			75	15	12 WNW	54	10	17	186
WP 1		Sr	2b	foreset			75	15	12 WNW	50	5	20	200
WP 1	92.0	St	2a	foreset			75	15	12 WNW	60	4	10	204
WP 1	94.0	St	2a	foreset	75	14	75	15	12 WNW	50	4	20	202
WP 1	96.5	Sh	2b	parting lineation			75	15	12 WNW	104-284			
WP 1	97.0	St	2d	foreset			75	15	12 WNW	90	358	21	3
WP 1	100.0	St	2a	Small scale forset			75	15	12 WNW	50	2	20	207
WP 1		Sp	2a	foreset			75	15	12 WNW	75	356	9	321
WP 1		Sp	2a	foreset			75	15	12 WNW	76	352	13	316
WP 1	123.0	St	2a	foreset	75	15	75	15	12 WNW	88	0	18	7
WP 1	163.7	Sp	2a	foreset			75	15	12 WNW	54	16	19	172
WP 1		Sp	2a	foreset			75	15	12 WNW	65	15	11	141
WP 1	168.0	St	1	foreset	75	14	75	15	12 WNW	70	6	2	119
WP 1		St	1	foreset			75	15	12 WNW	75	354	11	316
WP 1		St	1	foreset			75	15	12 WNW	90	16	23	53
WP 1		St	1	foreset			75	15	12 WNW	75	18	14	93
WP 1		St	1	foreset			75	15	12 WNW	60	10	11	176
WP 1	173.0	St	1	foreset	76	10	75	15	12 WNW	90	350	24	344
WP 1	182.0	Sl p	2d	foreset	85	20	75	15	12 WNW	80	358	12	348
WP 1		Sl p	2d	foreset			75	15	12 WNW	70	352	12	289
WP 1	94.0	St	2a	foreset	75	14	75	15	12 WNW	60	0	14	226
WP 1		St	2a	foreset			75	15	12 WNW	60	8	13	196
WP 1		St	2a	foreset			75	15	12 WNW	50	6	23	204

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section	m	facies	FA	indicator type	bedding		averaged bedding			vector measurement		rotated vector	
					dip°	dip dir	dip°	dip dir.	plunge	dip°	dip dir.	dip°	dip dir./trend
WP 1	101.0	Sr	2c	foreset	75	12	75	15	12 WNW	88	10	15	39
WP 1		Sr	2c	foreset			75	15	12 WNW	80	18	14	85
WP 1	104.0	Sl		foreset			85	8	12 WNW	84	196	23	74
WP 1	239.0	Sl	2a	foreset			85	8	12 WNW	68	22	28	143
WP 1		Sl	2a	foreset			85	8	12 WNW	85	0	4	70
WP 1		Sl	2a	foreset			85	8	12 WNW	70	14	21	148
WP 1	241.0	Sp	2a	foreset	85	8	85	8	12 WNW	73	6	13	158
WP 1	258.0	St	2a	foreset	85	8	85	8	12 WNW	68	356	15	203
WP 1	262.0	Sl	2a	foreset			85	8	12 WNW	68	4	16	176
WP 2	1.0	Sl	1	foreset	20	38	20	38	9 WNW	18	30	3	266
WP 2		Sl	1	foreset			20	38	9 WNW	26	36	6	38
WP 2		Sl	1	foreset			20	38	9 WNW	16	22	6	263
WP 2		Sl	1	foreset			20	38	9 WNW	5	24	15	224
WP 2		Sl	1	foreset			20	38	9 WNW	21	35	1	17
WP 2	4.0	Sl	1	foreset	20	50	20	38	9 WNW	25	54	8	100
WP 2	5.0	Sl	1	foreset			20	38	9 WNW	15	9	9	267
WP 2		Sl	1	foreset			20	38	9 WNW	21	40	2	86
WP 2	13.5	Sl	1	foreset	10	68	13	44	9 WNW	8	18	7	256
WP 2		Sl	1	foreset	15	28	13	44	9 WNW	5	186	17	215
WP 2	23.0	Sl	1	foreset			26	4	9 WNW	16	40	16	148
WP 2		Sl	1	foreset			26	4	9 WNW	15	52	20	150
WP 2	25.0	Sl	1	foreset			26	4	9 WNW	36	40	22	86
WP 2	25.5	Sl	1	foreset			26	4	9 WNW	30	48	21	106
WP 2	29.5	Sl p	1	foreset	25	27	26	4	9 WNW	30	32	15	93
WP 2		Sl p	1	foreset	27	20	26	4	9 WNW	33	36	18	88
WP 2	31.5	St	1	foreset			26	4	9 WNW	25	0	1	236
WP 2		St	1	foreset			26	4	9 WNW	20	333	12	236
WP 2		St	1	foreset			26	4	9 WNW	20	332	13	237
WP 2	33.0	Sl	1	foreset			26	4	9 WNW	10	25	17	173
WP 2		Sl	1	foreset			26	4	9 WNW	20	10	6	161
WP 2	45.5	Sl p	1	foreset	30	36	27	24	9 WNW	15	56	16	175
WP 2		Sl p	1	foreset			27	24	9 WNW	18	60	16	164
WP 2	46.0	Sl p	1	foreset			27	24	9 WNW	20	48	12	160
WP 2	57.5	Sl p	1	foreset			27	24	9 WNW	28	10	6	305
WP 2		Sl p	1	foreset			27	24	9 WNW	28	366	7	299
WP 2	59.0	Sl p	1	foreset	30	2	27	24	9 WNW	14	357	15	229
WP 2		Sl p	1	foreset	20	34	27	24	9 WNW	15	18	12	211
WP 2	60.0	Sl p	1	foreset			27	24	9 WNW	30	14	5	335
SF	2.0	Sp	2a	foreset	35	28	35.7	31.4	0	40	15	20	204
SF		Sp	2a	foreset	40	28	35.7	31.4	0	40	12	23	206
SF	4.5	Sl	2b	foreset	35	40	35.7	31.4	0	35	40	6	59
SF		Sl	2b	foreset			35.7	31.4	0	34	30	5	194
SF	12.0	St	2a	foreset			35.7	31.4	0	40	12	23	206
SF	13.0	Sr	2b	foreset			35.7	31.4	0	30	62	27	29
SF		Sr	2b	foreset			35.7	31.4	0	33	46	11	39
SF	14.0	St	2a	foreset			35.7	31.4	0	60	12	25	197
SF		St	2a	foreset			35.7	31.4	0	50	16	20	196
SF	15.0	St	2a	foreset			35.7	31.4	0	45	26	11	179
SF		St	2a	foreset			35.7	31.4	0	45	22	15	190
SF		St	2a	foreset			35.7	31.4	0	48	4	31	209
SF		St	2a	foreset			35.7	31.4	0	48	8	27	206
SF	19.0	Sp	2a	foreset			35.7	31.4	0	50	35	11	129
SF		Sp	2a	foreset			35.7	31.4	0	48	38	11	111
SF	22.0	Sp	2a	foreset			35.7	31.4	0	30	358	37	211
SF	25.0	St	1	foreset			35.7	31.4	0	24	68	33	19
SF		St	1	foreset			35.7	31.4	0	20	48	15	357
SF	27.0	Sm	1	foreset	36	34	35.7	31.4	0	35	12	23	209
SF		Sm	1	foreset			35.7	31.4	0	15	46	15	341
SF	31.0	St	1	foreset			35.7	31.4	0	16	2	33	212
SF		St	1	foreset			35.7	31.4	0	17	350	45	208
SF	38.0	Sp	2a	foreset			35.7	31.4	0	10	12	24	221
SF	38.5	Sp	2a	foreset			35.7	31.4	0	35	10	25	209
SF		Sp	2a	foreset			35.7	31.4	0	16	348	47	207
SF		Sp	2b	foreset			35.7	31.4	0	20	42	10	342
SF	40.5	Sr a	2b	foreset			35.7	31.4	0	50	14	22	199
SF		Sr a	2b	foreset			35.7	31.4	0	50	32	11	145
SF	44.5	St	1	foreset			35.7	31.4	0	24	14	21	216
SF	46.0	St	1	foreset			35.7	31.4	0	30	50	15	28
SF	47.5	St	2a	foreset			35.7	31.4	0	34	20	15	207
SF		St	2a	foreset			35.7	31.4	0	24	36	4	312
SF	49.0	St	2b	foreset			35.7	31.4	0	20	4	31	212
SF	49.8	St	2b	foreset	40	28	35.7	31.4	0	28	70	35	26
SF	50.8	St	2b	foreset			35.7	31.4	0	28	55	20	24
SF	51.5	Sp	2a	foreset			35.7	31.4	0	32	20	15	210
SF		Sp	2a	foreset			35.7	31.4	0	32	354	41	211
SF	56.0	Sl	2b	foreset			35.7	31.4	0	32	47	12	35
SF		Sl	2b	foreset			35.7	31.4	0	32	32	3	201
SF	58.0	St	2a	foreset			35.7	31.4	0	36	10	25	209
SF		St	2a	foreset			35.7	31.4	0	30	20	15	212
SF		St	2a	foreset			35.7	31.4	0	30	348	47	211

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section	m	facies	FA	indicator type	bedding		averaged bedding			vector measurement		rotated vector	
					dip°	dip dir	dip°	dip dir.	plunge	dip°	dip dir.	dip°	dip dir./trend
SF	60.0	St	2b	foreset			35.7	31.4	0	52	8	28	205
SF		St	2b	foreset			35.7	31.4	0	50	6	29	207
SF	60.2	St	2b	foreset			35.7	31.4	0	45	8	27	207
SF		St	2b	foreset			35.7	31.4	0	44	10	25	206
SF	67.5	Sp	2a	foreset			35.7	31.4	0	44	2	33	210
SF	75.0	Sp	2b	foreset			35.7	31.4	0	25	34	4	282
SF		Sp	2b	foreset			35.7	31.4	0	30	10	25	211
SF	80.0	Sp	2a	foreset			35.7	31.4	0	20	20	16	225
SF		Sp	2a	foreset			35.7	31.4	0	25	8	27	213
SF	81.0	Sp	2a	foreset			35.7	31.4	0	45	346	49	215
SF	81.8	Sp	2a	foreset	28	30	35.7	31.4	0	40	352	43	213
SF		Sp	2a	foreset			35.7	31.4	0	55	20	18	185
SF		Sp	2a	foreset			35.7	31.4	0	25	348	47	209
SF		Sp	2a	foreset			35.7	31.4	0	38	350	45	213
SF	89.0	Sl p	1	foreset			35.7	31.4	0	21	358	37	210
SF		Sl p	1	foreset			35.7	31.4	0	20	388	9	246
SF		Sl p	1	foreset			35.7	31.4	0	20	356	39	210
SF	91.0	Sl	1	foreset			35.7	31.4	0	28	0	35	211
SF		Sl	1	foreset			35.7	31.4	0	25	14	21	215
SF		Sl	1	foreset			35.7	31.4	0	25	352	43	210
SF		Sl	1	foreset			35.7	31.4	0	28	6	29	212
SF		Sl	1	foreset			35.7	31.4	0	25	356	39	210
SF	95.0	St	2a	foreset			35.7	31.4	0	26	348	47	210
SF		St	2a	foreset			35.7	31.4	0	22	24	12	229
SF	95.5	St	2a	foreset	38	40	35.7	31.4	0	9	342	52	203
SF		St	2a	foreset			35.7	31.4	0	34	348	47	212
SF	99.0	St	2a	foreset			35.7	31.4	0	50	4	31	208
SF	99.5	St	2a	foreset			35.7	31.4	0	52	18	19	191
SF		St	2a	foreset			35.7	31.4	0	50	26	13	172
SF		St	2a	foreset			35.7	31.4	0	50	20	17	189
SF		St	2a	foreset			35.7	31.4	0	44	8	27	207
SF	111.0	Sl	1	foreset			35.7	31.4	0	45	28	10	171
SF	115.0	St	2a	foreset	35	30	35.7	31.4	0	45	10	25	205
SF		St	2a	foreset			35.7	31.4	0	38	350	45	213
SF		St	2a	foreset			35.7	31.4	0	25	344	51	209
SF	118.5	St	2a	foreset			35.7	31.4	0	40	46	12	63
SF		St	2a	foreset			35.7	31.4	0	45	30	9	160
SF	124.0	St	1	foreset			35.7	31.4	0	25	34	4	282
SF	128.0	St	1	foreset			35.7	31.4	0	48	4	31	209
SF		St	1	foreset			35.7	31.4	0	50	20	17	189
SF	130.0	Sl p	1	foreset			35.7	31.4	0	30	348	47	211
SF		Sl p	1	foreset	35	20	35.7	31.4	0	42	358	37	212
SF		Sl p	1	foreset	35	36	35.7	31.4	0	30	6	29	211
RR	9.0	Sp	2a	foreset	75	210	70	211	0	70	220	8	303
RR	10.0	Sp	2a	foreset			70	211	0	85	220	17	242
RR		Sp	2a	foreset			70	211	0	70	230	18	304
RR	11.0	St	2a		75	210	70	211	0				
RR	13.0	St	2a	foreset			70	211	0	85	220	17	242
RR	14.0	St	2a	foreset			70	211	0	80	200	15	163
RR	19.0	St	2a	foreset	60	205	70	211	0	75	200	12	145
RR		St	2a	foreset			70	211	0	75	180	30	126
RR		St	2a	foreset			70	211	0	75	220	10	272
RR	23.0	St	2a	foreset	80	210	70	211	0	85	220	17	242
RR		St			70	235	70	211	0				
RR		St			65	210	70	211	0				
RR		St			60	220	70	211	0				
RR	33.0	St	2a	foreset	75	205	70	211	0	80	200	15	163
RR	34.0	St	2a	foreset	80	200	70	211	0	90	220	22	236
RR		St	2a	foreset			70	211	0	80	215	11	233
RR		St	2a	foreset			70	211	0	90	195	25	171
RR	39.0	St	2a	foreset			70	211	0	55	210	15	34
RR	45.0	St	2a	foreset	75	215	70	211	0	80	210	10	205
RR		St	2a	foreset			70	211	0	60	198	15	78
RR		St	2a	foreset			70	211	0	60	195	18	83
RR		St	2a	foreset			70	211	0	65	200	11	93
RR		St	2a	foreset			70	211	0	80	210	10	205
RR	45.6	St	2a	foreset			70	211	0	80	180	32	135
RR	53.0	St	2a	bed LBS			70	211	0	90	220	22	236
RR		St	2a	bed LBS			70	211	0	80	200	15	163
RR		St	2a	bed LBS			70	211	0	85	205	16	189
RR	54.0	Sl p	2a	foreset			70	211	0	75	210	5	200
RR	56.0	Sh	2a	foreset	80	210	70	211	0	80	210	10	205
RR	58.0	Sp	2a	foreset	75	210	70	211	0	70	220	8	303
RR		Sp	2a	foreset			70	211	0	75	215	6	249
RR	60.0	Sl p	2a	foreset			70	211	0	60	215	11	12
RR		Sl p	2a	foreset			70	211	0	75	215	6	249
RR		Sl p	2a	foreset			70	211	0	60	220	13	354
RR		Sl p	2a	foreset			70	211	0	60	210	10	36
RR	60.5	Sp	2a	foreset	65	210	70	211	0	70	210	1	121
RR	62.0	St	2a	foreset			70	211	0	75	205	8	161

section	m	facies	FA	indicator type	bedding		averaged bedding			vector measurement		rotated vector	
					dip°	dip dir	dip°	dip dir.	plunge	dip°	dip dir.	dip°	dip dir./trend
RR	63.0	Sl	2a	foreset			70	211	0	70	230	18	304
RR	64.0	Sl p	1	foreset			70	211	0	65	210	5	41
RR		Sl p	1	foreset			70	211	0	68	205	6	100
RR		Sl p	1	foreset			70	211	0	70	205	6	120
RR		Sp	1	foreset			70	211	0	85	209	15	203
RR		Sl p	1	foreset			70	211	0	85	220	17	242
RR		Sl p	1	foreset			70	211	0	70	210	1	121
RR	65.0	St	1	foreset	70	220	70	211	0	75	210	5	200
RR	67.0	Sp	1	foreset	65	190	70	211	0	90	200	23	181
RR	68.0	Sp	1	foreset	65	205	70	211	0	80	188	24	142
RR	70.0	St	1	foreset	60	220	70	211	0	90	205	21	194
RR		St	1	foreset			70	211	0	85	220	17	242
RR	72.0	Sp	1	foreset	65	220	70	211	0	75	210	5	200

Appendix 3 – Catalogue of trace fossils studied at the Albany Museum

Table 10-2: Table of abbreviations for the catalogue in Table 9-3

Section	Horizon	Reference
CH - Coombs Hill	L1, L2, L3, IO, IO sstn. (sandstone underlying IO)	Harris <i>et al.</i> 2021, <i>South African Journal of Science</i>
RR - Rabbit Ridge	LS (Lower Section), R1, R2	This work (Chapter 7)
WF - Waterloo Farm	MFL (Main Fish Lens), MFL sstn. (sandstone underlying MFL)	Scholze and Gess (2021) <i>Geobios</i>
HP - Howison's Poort	HP 1, HP 2	This work (Chapter 4)
FG - 'dam near Firglen turnoff'		This work (Chapter 3.1.2)

Table 10-3: Catalogue of studied trace fossils at Albany Museum

Formation	Section	Horizon	Museum catalogue no.	Identification(s)
Witpoort	CH	iO sstn.	AM 7740	<i>Scalichnus lepusaures</i> , Plug shaped burrows
			AM 7741	<i>Scalichnus phiale</i> , Plug shaped burrows
			AM 7746 a, b	<i>Ptychoplasma vagans</i>
			AM 7747	<i>Ptychoplasma vagans</i>
			AM 7748a	<i>Ptychoplasma vagans</i>
			AM 7748ai	<i>Scalichnus lepusaures</i>
			AM 7748b	<i>Ptychoplasma vagans</i>
			AM 7749ai	<i>Archaeonassa fossulata</i> , <i>Scalichnus phiale</i> , <i>Scalichnus lepusaures</i>
			AM 7749aia	<i>Archaeonassa fossulata</i> , <i>Ptychoplasma vagans</i> , <i>Scalichnus phiale</i>
			AM 7749b	<i>Archaeonassa fossulata</i> , Plug shaped burrows
			AM 7749d	<i>Scalichnus lepusaures</i>
			AM 7749f	<i>Scalichnus lepusaures</i>
		L3	AM 7750	Coprolite - spiral
		PP	AM 7744 a, b	<i>Ptychoplasma vagans</i>
			AM 7720 a, b	Coprolite - spiral
			AM 7721	Coprolite - spiral
			AM 7722	Coprolite - spiral
			AM 7723 a, b, c	Coprolite - spiral
			AM 7724	Coprolite - spiral
			AM 7725	Coprolite - spiral
			AM 7727 a, b	Coprolite - spiral
			AM 7728 a, b	Coprolite - spiral
			AM 7729 a, b	Coprolite - spiral

		AM 7730 a, b	Coprolite - spiral
		AM 7731 a, b	Coprolite - spiral
		AM 7732 a, b, c	Coprolite - spiral
		AM 7733 a, b	Coprolite - spiral
		AM 7734 a, b, c	Coprolite - spiral
		AM 7735 a, b	Coprolite - spiral
		AM 7737	Coprolite - spiral
		AM 7738	Coprolite - spiral
HP		AM 7643	<i>Zoophycos</i> cf. <i>brianteus</i>
		AM 7644 a, b	<i>Zoophycos</i> cf. <i>brianteus</i>
		AM 7645 a, b	<i>Zoophycos</i> cf. <i>brianteus</i> , <i>Planolites</i> ispp.
		AM 7646	<i>Zoophycos</i> cf. <i>brianteus</i>
FG		AM 7582	<i>Zoophycos</i> cf. <i>brianteus</i>
		AM 7604 a, b	<i>Zoophycos</i> cf. <i>brianteus</i>
		AM 7605	<i>Zoophycos</i> cf. <i>brianteus</i>
RR	R1	AM 7781 a, b	Coprolite - spiral
		AM 7782	<i>Nereites</i> ispp., <i>Palaeophycus tubularis</i> , <i>Phycodes</i> cf. <i>palmatus</i>
		AM 7783 a, b	<i>Zoophycos</i> cf. <i>brianteus</i>
		AM 7784 a, b	<i>Zoophycos</i> cf. <i>brianteus</i>
		AM 7785 a, b	<i>Nereites</i> ispp., Compound trace 1
		AM 7786 a, b	<i>Lingulichnus hamatus</i> , <i>Lingulichnus verticalis</i>
		AM 7787 a, b	<i>Lingulichnus verticalis</i> , Compound trace 1
		AM 7789 (bottom)	<i>Nereites</i> ispp., <i>Siphonichnidae</i> indet., Indet
		AM 7789 (top)	<i>Archaeonassa</i> cf. <i>fossulata</i> , <i>Siphonichnidae</i> indet., <i>Scalichnus phiale</i>
		AM 7791 a, b	<i>Lingulichnus verticalis</i>
		AM 7792	<i>Scalichnus phiale</i> , Compound trace 2, Compound trace 1
		AM 7794 a, b	<i>Palaeophycus tubularis</i>
		AM 7796	Compound trace 1
		AM 7797	Compound trace 1
		AM 7844	<i>Psammichnites implexus</i> , Indet.
	R2	AM 7802	<i>Palaeophycus tubularis</i> , <i>Planolites</i> ispp.
		AM 7803	<i>Planolites</i> ispp.
		AM 7806	<i>Lingulichnus verticalis</i>
		AM 7807	Coprolite - spiral
		AM 7808	Coprolite - spiral
		AM 7812 a, b	<i>Lingulichnus verticalis</i>
	LS	AM 7845	<i>Selenichnites rossendalensis</i>
WF	MFL	AM 18023	Coprolite - spiral
		AM 18024	Coprolite - spiral
		AM 18025 a, b	Coprolite - spiral
		AM 18026 a, b	Coprolite - spiral
		AM 18027 a, b	Coprolite - spiral
		AM 18028 a, b	Coprolite - spiral
		AM 18029	Coprolite - spiral
		AM 18030	Coprolite - spiral
		AM 18031 a, b	Coprolite - spiral

AM 18032 a, b	Coprolite - spiral
AM 18033 a, b	Coprolite - spiral
AM 18034	Coprolite - spiral
AM 18035 a, b	Coprolite - spiral
AM 18036 a, b	Coprolite - spiral
AM 18037 a, b	Coprolite - spiral
AM 18039	Diminutive plug shaped burrows
AM 18040 a, b	Diminutive plug shaped burrows
AM 18041	Diminutive plug shaped burrows
AM 18042	Diminutive plug shaped burrows
AM 18043	Micaceous backfill structures
AM 18044 a, b	Micaceous backfill structures
AM 18045	Micaceous backfill structures
AM 5266	<i>Spirophyton</i>
AM 5267	<i>Spirophyton</i>
AM 5292	<i>Spirophyton</i>
AM 5397	Coprolite - spiral
AM 5398	Coprolite - spiral
AM 5401 E	Coprolite - spiral
AM 5401 F	Coprolite - spiral
AM 5401 G	Coprolite - spiral
AM 5401 H	Coprolite - spiral
AM 5401 I	Coprolite - spiral
AM 5401 O	Coprolite - spiral
AM 5404	<i>Spirophyton</i>
AM 5489	Coprolite - spiral
AM 5723	Coprolite - spiral
AM 5724	Coprolite - spiral
AM 5725	Coprolite - dumbell shaped
AM 5735	Coprolite - oval
AM 7607 a, b, c	<i>Spirophyton</i>
AM 7608	<i>Spirophyton</i>
AM 7612 a, b	<i>Spirophyton</i>
AM 7615	<i>Undichna</i> ispp., Plug shaped burrows, Compound trace 1
AM 7616	<i>Undichna</i> ispp.
AM 7617	<i>Undichna</i> ispp.
AM 7618	<i>Undichna</i> ispp.
AM 7619	<i>Undichna</i> ispp.
AM 7621 a, b	Regurgitalite
AM 7622 a, b	Regurgitalite
AM 7623 a, b	Regurgitalite
AM 7629 a, b	Coprolite - spiral
AM 7630 a, b	Coprolite - spiral
AM 7631 a, b	Coprolite - spiral
AM 7632	Coprolite - spiral
AM 7633	Coprolite - spiral

			AM 7634 a, b	Coprolite - spiral
			AM 7635	<i>Spirophyton</i>
			AM 7636 a, b	Diminutive plug shaped burrows
			AM 7637	Coprolite - spiral
			AM 7638	Coprolite - spiral
			AM 7639	Coprolite - spiral
			AM 7642 (a + b)	Coprolite – spiral, Coprolite - oval
			AM 7652 a, b i-iii	<i>Spirophyton</i>
			AM 7653	<i>Spirophyton</i>
		MFL sstn	AM 5402 c, e, f, j, l, n, v, w, y, z	<i>Cruziana problematica</i> , <i>Rusophycus carbonarius</i>
			AM 5403	<i>Cruziana problematica</i> , <i>Rusophycus carbonarius</i>
			AM 5625	<i>Cochlichnus anguineus</i> , Indet.
			AM 7606 a, b	<i>Phycodes</i> cf. <i>palmatus</i> , Compound trace 1
			AM 7609 a, b	<i>Cruziana problematica</i> , <i>Rusophycus carbonarius</i>
			AM 7610	<i>Cruziana problematica</i> , <i>Phycodes</i> cf. <i>palmatus</i> , <i>Rusophycus carbonarius</i>
			AM 7611	<i>Cruziana problematica</i> , Plug shaped burrows
			AM 7625	<i>Cruziana problematica</i> , <i>Rusophycus carbonarius</i> , Compound trace 1
			AM 7626	<i>Cruziana problematica</i> , <i>Lockeia siliquaria</i> , Compound trace 1
			AM 18021	Coprolite – spiral, Coprolite - oval
			AM 7680 a, b	<i>Spirophyton</i>
			AM 7681 a, b, c	<i>Spirophyton</i>
		WF 1	AM 18038	<i>Ptychoplasma conica</i>
			AM 7647	<i>Planolites</i> ispp., <i>Ptychoplasma vagans</i>
			AM 7658a	<i>Archaeonassa fossulata</i>
			AM 7661	<i>Archaeonassa fossulata</i>
			AM 7664 a, b	<i>Ptychoplasma vagans</i>
			AM 7670	Compound trace 1
			AM 7672	<i>Lockeia siliquaria</i>
			AM 7673	Compound trace 1
			AM 7674	Compound trace 1
			AM 7675	Compound trace 1
			AM 7683 i, ii	<i>Lockeia siliquaria</i>
		HP 2	AM 7671	<i>Gordia marina</i>
Weltevrede	HP 1			

Appendix 4 – Supplementary material to Chapter 7 (as published online by Harris and Gess 2022, *Palaios*)

APPENDIX 4A– Catalogue of lingulid brachiopod fossils

Locality	Lingulid specimens		Dimensions (mm)		W:L	Pedicel or brachial valve	
	Catalogue no.	Field no.	Length	Width			
R1 locality, Rabbit Ridge, Eastern Cape (33°17'47.6"S 26°44'29.5"))							
	AM 7814	R(1) 008 a	17.7	14.5	0.82	?	
		b	16.8	12.5	0.74	?	
	AM 7815	R(1) 009 a	21.7	15.8	0.73	Articulated	
		b	20.6	14.0	0.68	P	
		R(1) 009 b	18.1	14.3	0.79	P	
		b	20.0	13.8	0.69	B	
		c	20.9	14.1	0.67	B	
	AM 7816	R(1) 010	22.3	16.1	0.72	P	
		b	22.3	18.8	0.84	B	
		c	24.2	15.4	0.64	P	
		d	23.0	15.2	0.66	B	
		e	12.9	9.6	0.74	P	
	AM 7817	R(1) 011	21.7	12.8	0.59	P	
		b	16.9	13.3	0.79	?	
		c	15.6	11.4	0.73	B	
		d	18.8	9.6	0.51	B	
	AM 7818	R(1) 012	24.3	15.0	0.62	B	
		b	22.7	17.4	0.77	B	
		c	13.9	8.4	0.60	?	
	AM 7813 b	R(1) 013	21.0	15.9	0.76	P	
	AM 7813 c	R(1) 014	19.9	12.3	0.62	P	
	AM 7813 d	R(1) 015	23.9	17.7	0.74	B	
	AM 7813 e	R(1) 016	19.0	11.9	0.63	?	
	AM 7813 f	R(1) 015 dup	19.4	12.2	0.63	B	
	AM 7813 g	R(1) 017 dup	19.2	12.2	0.64	B	
	AM 7813 h	R(1) 018 a	20.6	14.5	0.70	B	
	AM 7813 i	R(1) 019	19.9	13.3	0.67	B	
	AM 7813 j	R(1) 021	18.1	13.4	0.74	B	
	AM 7813 k	R(1) 022	20.7	14.1	0.68	P	
	AM 7813 l	R(1) 023	20.5	15.7	0.77	B	
	AM 7813 m	R(1) 024	19.8	16.1	0.81	P	
	AM 7813 n	R(1) 025	21.5	13.3	0.62	P	
	AM 7813 o	R(1) 044	24.0	17.9	0.75	?	
	AM 7819	R(1) 045 a	19.2	13.0	0.68	?	
		b	14.9	12.9	0.87	B	
	AM 7813 p	R(1) 047	19.0	14.0	0.74	B	
	AM 7820	R(1) 048 a	19.5	15.0	0.77	B	
		b	24.8	15.5	0.63	P	
	AM 7813 q	R(1) 049	23.4	16.3	0.70	P	
	AM 7813 r	R(1) 057	25.5	20.0	0.78	B	
	AM 7821 a	SS 001	19.6	13.7	0.70	P	
	AM 7821 b	SS 002	15.8	10.6	0.67	?	
	AM 7821 c	SS 003	19.8	14.4	0.73	B	
	AM 7822	SS 004 a	24.5	14.8	0.60	P	
		b	26.0	16.4	0.63	?	
		c	19.3	14.7	0.76	B	
		d	22.0	18.3	0.83	P	
		e	23.9	17.5	0.73	B	
	AM 7821 d	SS 005	16.0	10.5	0.66	?	
	AM 7821 e	SS 006	19.6	14.9	0.76	B	
	AM 7821 f	SS 007	20.2	12.2	0.60	?	
	AM 7821 g	R1-SS 020	23.1	18.8	0.81	?	
	AM 7841	R1-SS 021 a	22.5	15.3	0.68	P	
		b	22.9	14.1	0.62	?	
	AM 7821 h	R1-SS-022	21.3	13.1	0.62	?	
	AM 7821 i	R1-SS 028	14.5	10.1	0.70	?	
	AM 7821 j	R1-SS 029	16.8	11.3	0.67	P	
	AM 7821 k	R1-SS 030	21.5	15.6	0.73	B	
	AM 7821 l	R1-SS 031	19.5	12.7	0.65	P	
	AM 7821 m	R1-SS 032	21.2	13.8	0.65	?	
	AM 7821 n	R1-SS 033	24.2	13.7	0.57	B	

Locality	Lingulid specimens			Dimensions (mm)		W:L		Pedicle or bra
	Catalogue no.	Field no.		Length	Width			
R2 locality, Rabbit Ridge, Eastern Cape (33°17'18"S 26°43'25.6"E)								
	AM 7822	R(2) 002	a*	19.0	12.5	0.66		?
			c	20.2	13.2	0.65		P
	AM 7823 a	R(2) 009		20.0	12.8	0.64		B
	Am 7823 b	R(2) 013 a+b		21.2	14.5	0.68		?
	AM 7823 c	R(2) 014		19.2	14.3	0.74		?
	AM 7823 d	R(2) 018		16.5	11.4	0.69		B
	AM 7824	R(2) 020	a	21.6	14.6	0.68		P
			b	22.2	14.6	0.66		?
			c	17.4	13.1	0.75		P
			d	20.2	13.7	0.68		B
			e	21.9	14.1	0.64		B
			f	22.5	16.4	0.73		B
	AM 7825	R(2) 021	a	17.7	11.3	0.64		P
			b	18.6	12.5	0.67		B
	AM 7823 e	R(2) 024		20.7	14.8	0.71		B
	AM 7826	R(2) 037	a	21.7	14.8	0.68		B
			b	20.7	14.0	0.68		B
			d	18.9	13.5	0.71		P
	AM 7827	R(2) 040	a	20.5	14.2	0.69		B
			b	19.6	13.7	0.70		B
	AM 7828	R(2) 042	a	7.0	4.7	0.67		P
			b	6.4	4.1	0.64		?
	AM 7823 f	R(2) 043		17.8	11.9	0.67		P
	AM 7823 g	R(2) 044		14.2	9.5	0.67		B
	AM 7823 h	R(2) 045		20.3	13.7	0.67		P
	AM 7823 i	R(2) 049		19.6	13.2	0.67		P
	AM 7823 j	R(2) 051		24.6	16.4	0.67		P
	AM 7829	R(2) 052	a	21.5	14.8	0.69		P
			b	21.0	14.5	0.69		B
	AM 7823 k	R(2) 053		21.2	16.2	0.76		?
	AM 7823 l	R(2) 054		17.5	13.7	0.78		B
	AM 7823 m	R(2) 055		22.0	14.2	0.65		B
	AM 7823 n	R(2) 056		23.1	16.5	0.71		B
	AM 7823 o	R(2) 057		18.9	13.3	0.70		?
	AM 7823 p	R(2) 058		18.5	12.5	0.68		?
	AM 7823 q	R(2) 057 dupl		17.5	13.1	0.75		?
	AM 7823 r	R(2) 059		20.8	14.2	0.68		P
	AM 7823 s	R(2) 060		18.6	13.1	0.70		P
	AM 7823 t	R(2) 061		21.1	14.5	0.69		B
	AM 7823 u	R(2) 063		22.3	15.7	0.70		B
	AM 7823 v (a + b)	R(2) 065		21.2	13.3	0.63		P
	AM 7823 w	R(2) 066		20.8	13.0	0.63		P
	AM 7830	R(2) 068	a	19.2	13.6	0.71		B
			b	21.7	15.7	0.72		B
	AM 7823 x	R(2) 071		15.7	11.9	0.76		?
	AM 7823 y	R(2) 072		20.2	13.7	0.68		?
	AM 7823 z	R(2) 073	a	16.6	11.0	0.66		?
	AM 7831 a	R(2) 074		15.1	10.6	0.70		P
	AM 7831 ba	R(2) 075		21.6	14.5	0.67		P
	AM 7831 c	R(2) 076		16.2	10.9	0.67		?
	AM 7832	R(2) 078	a	17.8	12.1	0.68		B
			b	19.5	14.5	0.74		B
			c	21.2	13.3	0.63		P
	AM 7833	R(2) 079	a	24.2	16.3	0.67		B
			b	16.2	10.8	0.67		P
	Am 7834	R(2) 080	a	19.7	13.3	0.68		P
			b	20.3	14.1	0.69		B
	AM 7831 d	R(2) 082		15.4	9.2	0.60		P
	AM 7831 e	R(2) 083		19.0	13.5	0.71		B
	AM 7831 f	R(2) 085		14.4	8.8	0.61		P
	AM 7831 g	R(2) 086		19.4	12.8	0.66		P
	AM 7831 h	R(2) 093		7.4	5.0	0.68		?

Appendices

Lingulid brachiopods from the Oosthuizen collection, Iziko Museum, Cape Town									
Locality	Catalogue number		Length (cm)	Width (cm)	W:L		Identification	Identified by	Formation
Langvlei, Porterville, Western Cape, (33°01' S 19°03'E)									
	SAM-PB-16692		2.2	1.5	0.66		<i>Lingula</i> sp.	Oosthuizen	Disa Member, Cedarberg Formation, Table Mountain Group
	SAM-PB-16707		1.3	1.0	0.73		<i>Lingula</i> sp.	"	Disa Member, Cedarberg Formation, Table Mountain Group
Grootriver, Ceres, Western Cape, (32°38'S 19°24.5'E)									
	SAM-PB-14920		2.4	1.7	0.71		<i>Lingula lamella</i> Clarke	"	Gydo Fm., Bokkeveld Group ("Bokkeveld 1st shale")
	SAM-PB-14880		1.8	1.1	0.58		<i>Lingula keideli</i> Clarke	"	Gydo Fm., Bokkeveld Group ("Bokkeveld 1st shale")
Gamka Poort									
	SAM-PB-14973		1.4	0.8	0.57		<i>Lingula</i> sp.	"	? Fm, Bokkeveld Group
Zwaarmoed Pass, Ceres (33°21.5'S 19°30.5'E)									
	SAM-PB-14806		1.3	0.8	0.62		<i>Lingula lepta</i>	"	Voorstehoek Fm., Bokkeveld Group ("Bokkeveld 2nd shale")
	SAM-PB-15053		2.0	1.4	0.68		<i>Lingula lepta</i>	"	Voorstehoek Fm., Bokkeveld Group ("Bokkeveld 2nd shale")
	SAM-PB-15067		1.6	1.0	0.65		<i>Lingula keideli</i>	"	Voorstehoek Fm., Bokkeveld Group ("Bokkeveld 2nd shale")
Matroosberg, Worcester, Western Cape, (33°30'S 19°49'E)									
	SAM-PB-14932		2.5	1.4	0.57		<i>Lingula lepta</i>	"	Voorstehoek Fm., Bokkeveld Group ("Bokkeveld 2nd shale")
	SAM-PB-14802		1.7	1.0	0.59		<i>Lingula lepta</i>	"	Voorstehoek Fm., Bokkeveld Group ("Bokkeveld 2nd shale")
Theronsberg Pass, Ceres, 19 31.5' 33 14.5									
	SAM-PB-15052		2.7	1.3	0.48		<i>Lingula lepta</i>	"	Bokkeveld group 3rd shale
Tafelberg, Ceres, (32°56.75'S 19°21.75'E)									
	SAM-PB-14862		1.1	0.8	0.73		<i>Lingula</i> sp. b	"	Tra-Tra Fm., Bokkeveld Group ("Bokkeveld 4th shale")
	SAM-PB-14858		1.3	0.9	0.65		<i>Lingula</i> sp. b	"	Tra-Tra Fm., Bokkeveld Group ("Bokkeveld 4th shale")
	SAM-PB-14865		1.5	0.9	0.57		<i>Lingula</i> sp. b	"	Tra-Tra Fm., Bokkeveld Group ("Bokkeveld 4th shale")
	SAM-PB-14857						<i>Lingula</i> sp. b	"	Tra-Tra Fm., Bokkeveld Group ("Bokkeveld 4th shale")
	SAM-PB-14993		0.7	0.5	0.64		<i>Lingula</i> sp. b	"	Tra-Tra Fm., Bokkeveld Group ("Bokkeveld 4th shale")
	SAM-PB-14863		1.2	0.9	0.75		<i>Lingula</i> sp. b	"	Tra-Tra Fm., Bokkeveld Group ("Bokkeveld 4th shale")
Touws River (see Broquet et al.1983)									
	F85		1.9	1.3	0.68		<i>Lingula lepta</i>	"	Wagen Drift Fm. Witteberg Group
	F86		1.5	1.0	0.67		<i>Lingula lepta</i>	"	Wagen Drift Fm. Witteberg Group
	F14		1.7	1.1	0.67		<i>Lingula lepta</i>	"	Wagen Drift Fm. Witteberg Group
	Q11		2.1	1.4	0.64		<i>Lingula lepta</i>	"	Wagen Drift Fm. Witteberg Group

Appendix 4B– Description and Identity of Lingulid Fossils

Phylum BRACHIOPODA

Class LINGULATA Gorjansky & Popov, 1985

Order LINGULIDA Waagen, 1885

Family LINGULIDAE Menke, 1828

Genus *Dignomia* Hall, 1871

Species *Dignomia* (*Lingula*) cf. *lepta* Clarke, 1912

Material. — AM 7813 – AM 7839 (individual valves designated lower case alphabetical letters)

Locality.— Rabbit Ridge, Eastern Cape, South Africa, occurrences extending over three kilometres at one stratigraphic horizon.

Stratigraphic Horizon.— lower Witpoort Formation, Witteberg Group

Age.— early Famennian (Late Devonian)

Occurrence of D. lepta.— ‘*Lingula*’ *lepta* is the only lingulid species reported from the Witteberg Group (Hiller 1990) and occurs within underlying Early to Middle Devonian strata of the Bokkeveld Group (Cape Supergroup) alongside three other species assigned to ‘*Lingula*’ (Oosthuizen 1984). Very similar occurrences of this species in the Weltevrede Formation (lower Witteberg Group) and the Jaguariaiva Member (Ponta Grossa Fm, Paraná basin) comprising lingulids associated with orbiculoid brachiopods and various bivalves (particularly infaunal nucloids) are interpreted as representing nearshore or backshore habitats (Copper 1977; Hiller 1990; but see Horodyski *et al.* 2019).

The species is recorded at several localities in the upper Wagen Drift Formation in the Western Cape (Givetian/Frasnian, lower Weltevrede Subgroup), commonly associated with articulate brachiopod *Tropidoleptus*, a cosmopolitan Devonian taxon, and infaunal bivalves (Boucot *et al.* 1983). A lingulid-orbiculoid brachiopod association, with *Tropidoleptus*, various bivalve genera (*Palaeoneilo* sp., *Sanguinolites* sp., *Modiomorpha?* sp.), a gastropod

(*Plectonotus*) and a homalonotid trilobite (*Trimerus?*) are recorded in the uppermost Weltevrede Formation in the Eastern Cape (Hiller 1990).

In a review of the genus, Emig and Herrera (2006) concluded that *Dignomia* ranges from Middle Ordovician to Middle Devonian strata. This conclusion was likely based on the South American fossil record, but the Rabbit Ridge occurrences indicate that the genus persisted until the Famennian. This may be due to preservational biases, as Late Devonian strata from high-latitude Gondwana are rarely preserved. The Witteberg Group may exhibit the only complete Late Devonian succession from the palaeo-region which has not been wholly or partially removed by glacial action (de Vargas *et al.* 2020).

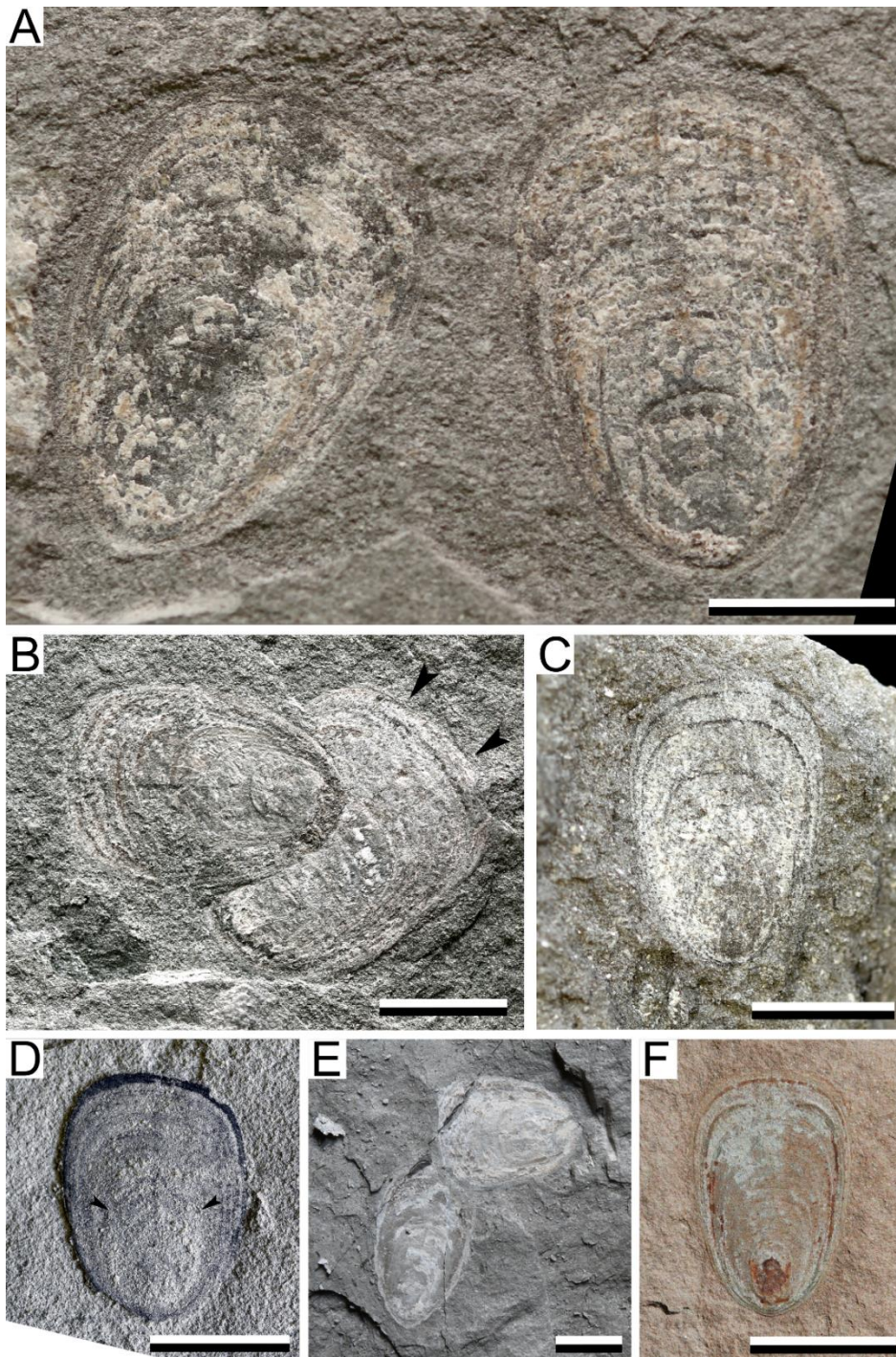


Figure 10.1: SUPPLEMENTAL FIG. 1, *Dignomia* cf. *lepta* from the Late Devonian of South Africa (all scale bars represent 1 cm). A) Similarly sized dorsal (right) and ventral (left) valves (AM 7829b). B) Overlay of a ventral (left) and a dorsal valve (AM 7826), note the mediolateral striations (arrowed). C) A dorsal valve (AM 7831a). D) A dorsal valve with possible anterior adductor muscle scars arrowed (AM 7831e). E) A narrow elongate ventral valve and a wider, spatulate ventral valve (AM 7840a). F) A ventral valve (AM 4782; ‘*Lingula*’ *lepta*) from a locality in the Wagen Drift Formation (Lower Witteberg Group) near Touws River (Western Cape), note the clear mediolateral striations on the valve.

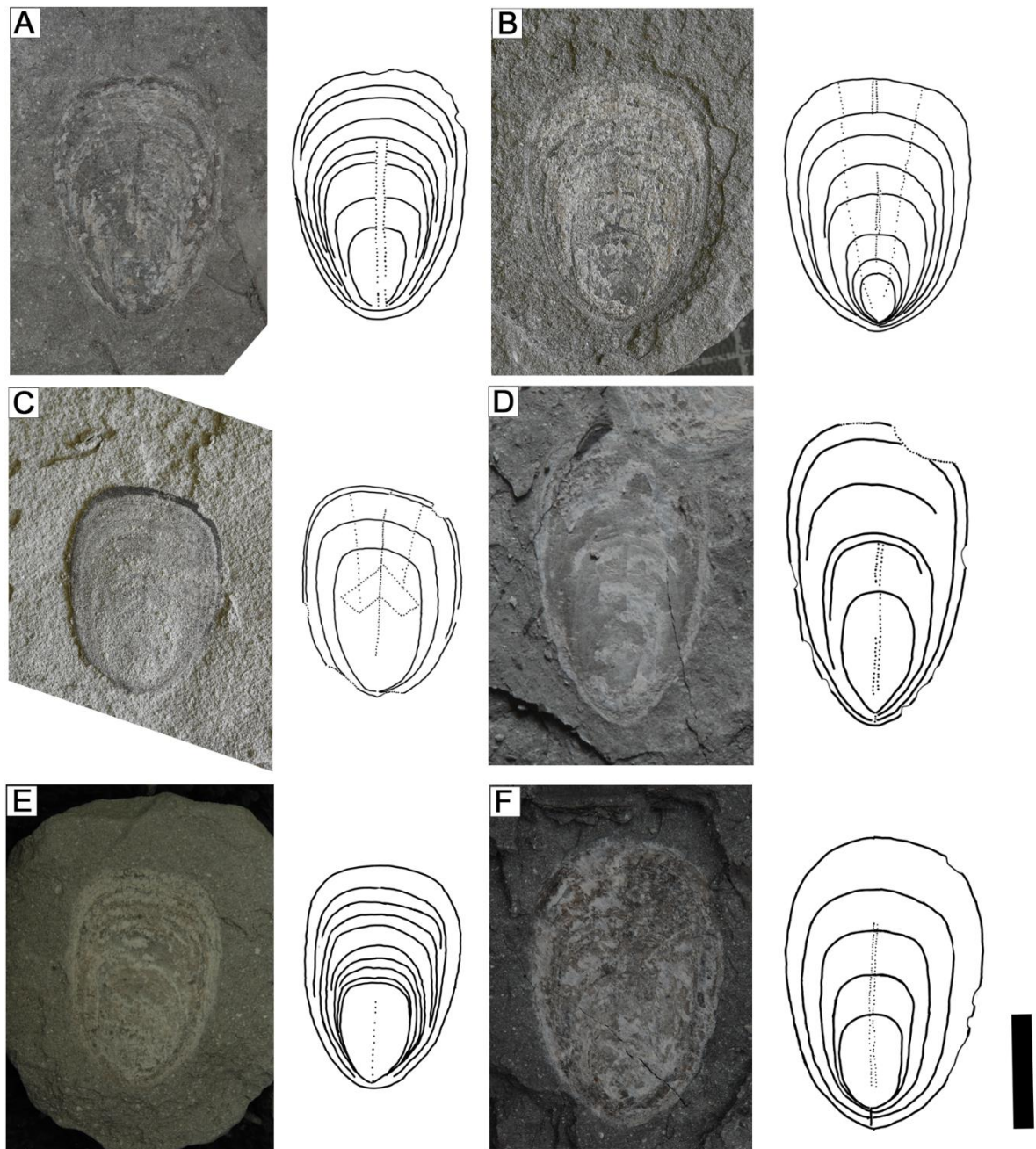


Figure 10.2: SUPPLEMENTAL FIG. 2— A selection of *Dignomia cf. lepta* shells from the Rabbit horizon with corresponding line drawings (all to same scale: scale bar = 1 cm). A) AM 7841a, B) AM 7829b, C) AM 7831e, D) AM 7840a, E) AM 7823p, F) AM 7833.

Description.— Shells are bilaterally symmetrical and vary in shape from ovoid to spatulate and the population exhibits clines of variation between end-members (Supp. Fig. 1

A–F). Spatulate forms have a nearly straight anterior margin and subparallel lateral margins, whereas ovoid forms have a semicircular anterior margin with curved lateral margins. Valves are greatest in width usually between half and two-thirds of their length, but sometimes more anteriorly (in spatulate forms). Lateral margins form an open curve, tapering to the umbo, which forms an obtuse angle in ventral (pedicle) valves, but is round in dorsal (brachial) valves. Dorsal and ventral valves commonly occur together on bedding surfaces. Adjacent, similarly sized dorsal and ventral valves indicate those of the same individual (Supp. Fig. 1A).

Shells are ornamented by up to 10 clear radial growth lines (Supp. Fig. 2), which are probably annual growth increments. A medial lineation is visible extending from the umbo to the anterior margin in many valves though it is not apparent or only partially so in others. This striation is most apparent on the anterior half to two thirds of dorsal valves whereas in pedicle valves it is strongly developed proximal to the umbo. Lineations forming a v-shape from the umbo to the anterior margins are faintly discernible in some specimens, and are usually clearest near to the anterior valve margins (Supp. Fig. 1B, C). Comparison with Devonian lingulids (*'Lingula' lepta*) from stratigraphically lower in the Cape Supergroup shows that these markings are a common feature (Supp. Fig. 1F).

Common faint traces of muscle scarring are visible on the specimens, including possible anterior adductor muscle scars in one specimen preserved in fine sandstone (Supp. Fig. 1C), however, in the majority of specimens, including those preserved in mudstone, muscle scars are poorly defined due to phyllosilicate mineralisation of the organic material comprising the valves.

The shells range in length from 6.4 mm to 26 mm (although 97% of 123 collected individuals are greater than 13 mm in length) and in width from 4.1 mm to 20 mm. The mean W:L for the sample is 0.69 (n=123) and the range in W:L for the sample is 0.57–0.84. Within

a collected sample of 123 bedding parallel valves, 34% are identified as ventral valves, 40.1% as dorsal valves and 24.4% are uncertain (see Supp. Mat. A). Only three bedding parallel specimens in the collection remain articulated, in ‘scissor position’, but the number is likely higher because of the potential of unrecognised valve superimposition.

Remarks.— The lingulid fossils described here are attributed to *Dignomia* based on the presence of distinctive median and mediolateral lineations (originally external grooves) on the shells (Emig and Herrera 2006). Although partially obscured by compression and remineralisation in much of the Rabbit Ridge sample, the median groove (internal septum) is evident in almost all specimens and mediolateral grooves in several. Although suggested to bear an asymmetrical muscle system characteristic of the family Lingulidae (Emig and Herrera 2006), muscle scars of Devonian *Dignomia* are unknown.

We tentatively compare these fossils to *Dignomia leptota* Clarke 1913, due to their relatively broad shells and overall shape, which are consistent with the species (Emig and Herrera 2006). Range in width to length ratios encompasses those reported for *D. leptota*, *D. alveata* and *D. subalveata*, (recorded from very small samples), but distinction between these species is unclear (Emig and Herrera 2006). The W:L ratio reported for these species (0.72–0.81) in South America falls within the range exhibited by the Rabbit Ridge population. Thus, a review of Devonian *Dignomia* is required for confident species-level identification, which may test whether *D. leptota* is a junior synonym of *D. subalveata* Kayser 1897, as was suggested previously (Castellaro 1966).

Rabbit Ridge *Dignomia* are clearly distinguished from Ordovician species *D. munsterii*, *D. lineata* and *D. submarginata* on the basis of W:L ratio, all of which have relatively long, narrow, linguliform shells with W:L ratios between 0.45 and 0.55 which does not overlap with the ranges recorded from Rabbit Ridge, but share the characteristic internal ridges of the shell that unite the genus (Emig and Herrera 2006).

Dignomia (Lingula) lepta was originally described from late Early to Middle Devonian strata of the Paraná basin in Brazil, alongside four other species attributed to the extant genus *Lingula* (Clarke 1913). Clarke separated the new species from *Dignomia* based on having a thinner shell, and included *Dignomia subalveata* in synonymy with *lepta*. The species were separated based on shell shape and size. '*Lingula*' *lepta* was described as having an elongate subquadrate shape with rounded dorsal and ventral angles. The divergent 'muscle scars' and faint median septum recorded by Clarke correspond to the median and mediolateral striations visible on the Rabbit Ridge specimens, features that are visible in Clarke's (1913, Plate XXVI) illustrations of the species. No data on size range or W:L was recorded in the description of *lepta*. Amongst Clarke's species '*Lingula*' *lamella* is distinguished from *lepta* based on its larger size (up to 33 mm long) and more quadrate outline. '*Lingula*' *keidelli* is a spatulate form with a subacute umbo. '*L*' *scalprum* is a small, narrow form with a distinct median ridge in the umbonal portion of the shell. Muscle scars and W:L ratios are unknown for these species.

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