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Distinct petroacoustic signature in Heterozoan and Photozoan carbonates resulting from combined depositional and diagenetic processes 3

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19 Abstract

The acoustic properties of heterozoan and photozoan carbonates are compared and 20 interpreted in the light of their distinct patterns of diagenetic evolution. The combined role of 21 depositional and diagenetic processes on the petroacoustic signatures of both heterozoan and 22 photozoan carbonates has been assessed on the basis of a multi-disciplinary and multi-scale 23 study of Oligo-Miocene limestone outcrops from the Cape Range peninsula (NW Australia). 24 This study integrates a detailed diagenetic analysis (using sedimentology, petrography and 25 stable isotopes geochemistry), as well as physical properties characterization (through 26 27 acoustic velocity laboratory and porosity measurements, petroacoustic modelling and 3D CTscan imaging). The three formations studied are characterized by distinct patterns of 28 diagenetic and petroacoustic evolution: heterozoan, non-tropical carbonates from the Mandu 29 and Tulki formations are characterized by significant development of microporosity and 30

pressure-solution compaction features which led to decreasing the pore network stiffness. Photozoan, sub-tropical to tropical carbonates from the Trealla formation are dominated by moldic, vuggy and partially cemented intergranular pores which tends to increase the pore network stiffness. The distinct petroacoustic signatures of heterozoan and photozoan limestones have been shown to be mainly related to (1) differences in compaction during burial resulting from an emersion phase and (2) differences in the mineralogy of dominant carbonate producers.

Temperate High-Mg calcite dominated carbonate typically poorly cemented allowed for 38 39 greater compaction degree, reducing the ratio of intergranular macropore. Furthermore, these dominant calcite producers (large benthic foraminifers, red algae...) from the heterozoan 40 Mandu and Tulki formations favoured the development of intragranular microporosity 41 42 whereas less-stable aragonite dominated biota from the photozoan Trealla Limestone favoured the development of molds and vugs, during early meteoric diagenesis. Finally, 43 44 petroacoustic modelling approaches such as α_K inversion have proved to be efficient for quantifying the impact of diagenetic processes and associated pore types on acoustic 45 46 properties. These results, emphasizing the strong variability in depositional and diagenetic 47 features for distinct carbonate fabrics could have interesting implications in terms of reservoir properties 3D modelling and upscaling approaches. 48

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50 Keywords: carbonates; diagenesis; ultrasonic velocities; petroacoustic modelling;
 51 heterozoan; photozoan; Oligo-Miocene; NW-Australia; outcrop analogue

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56 **<u>1. Introduction</u>**

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Acoustic properties of carbonate reservoirs at core scale have been shown to be largely 58 59 controlled by pore space architecture determined by depositional and diagenetic processes, including mineral transformations (e.g. Anselmetti and Eberli, 1993; Eberli et al., 2003; 60 61 Adelinet et al., 2019; Bailly et al., 2019a). In pure calcitic limestone reservoirs, porosity and 62 pore geometry are the main factors controlling P and S-wave velocities (e.g. Weger, 2009; Hairabian et al., 2014; Soete et al., 2015; Abdlmutalib et al., 2019). The variety of diagenetic 63 processes (cementation, dissolution, neomorphism, etc) affecting carbonate rocks commonly 64 result in a wide range of pore types, porosity and wave velocity as well as in a wide scattering 65 66 of values in velocity-porosity relationships (e.g. Verwer et al., 2008; Brigaud et al., 2010).

In carbonates rocks, pore type is an important factor controlling reservoir properties and fluid 67 flow at the matrix scale (e.g. Borgomano et al., 2013). In particular, this explains the 68 69 significant scatter observed in porosity/permeability relationships existing in different 70 geological contexts (Ehrenberg et al., 2006; Lønøy, 2006). Because of the high chemical reactivity of carbonate minerals (e.g. Ehrenberg and Nadeau 2005; Richard, 2014), the impact 71 of diagenetic transformations on porosity and pore space architecture is considerable. 72 73 Assessing such an impact on velocity-porosity relationships is thus a major issue for the prediction of reservoir properties from remote data such as well logs and seismics 74 (Anselmetti and Eberli, 1999; Fournier et al., 2018; Jafarian et al., 2018; Teillet et al., 2020) 75 76 and to achieve valid porosity-permeability transforms, which are of prime importance in the modelling of carbonate hydrocarbon systems. Most of the published studies dealing with 77 velocity-porosity relationships in purely calcitic (%CaCO₃ > 95 %) marine carbonate 78 reservoirs focused on tropical photozoan shallow-water carbonates (e.g. Anselmetti and 79

80 Eberli, 1993; Verwer et al., 2008; Brigaud et al., 2010; Fournier et al., 2014) or deep-water carbonates (Røgen et al., 2004; Hairabian et al., 2014). However, shallow-water non-tropical 81 heterozoan carbonates (sensu James, 1997) have received less attention, with regard to their 82 83 petroacoustic properties. Heterozoan carbonates are different from photozoan carbonates in terms of the carbonate producing biota, mineralogy, depositional profile (lack of barrier reef, 84 variable slope ramp, etc) and carbonate productivity (e.g. James, 1997; Wilson, 2002; 85 Schlager, 2003, Michel et al., 2018). By considering the differential stability between 86 aragonite, high Mg- and low Mg-calcite, distinct patterns of diagenetic evolutions and 87 88 petroacoustic signatures are therefore expected between heterozoan and photozoan 89 carbonates while interacting with similar fluids.

90 The present study is based on the sedimentological, petrographical, diagenetic and petro-91 acoustic analysis of the Oligo-Miocene shallow-water heterozoan to photozoan carbonates 92 from the Cape Range Group outcropping in the Northern Carnarvon Basin (NCB), in the most proximal part of the North West Shelf of Australia (Fig. 1A) (Hocking, 1987; Apthorpe, 93 94 1988; Collins et al., 2006; Riera et al., 2019). Offshore, age-equivalent carbonate sequences represent a thick part of the overburden, directly overlying the Mesozoic producing 95 hydrocarbon reservoirs. They constitute a major target for regional hydrocarbon exploration 96 and production (Wallace et al. 2003). The overburden exhibits strong vertical and lateral 97 velocity variations believed to originate from depositional and diagenetic heterogeneities. It 98 99 causes considerable seismic imaging and interpretation challenges, thus increasing uncertainties in regional reservoir and structural trap delineation, as well as major drilling 100 hazards (Wallace et al., 2003; Shragge et al., 2019). However, the depositional architecture 101 and diagenetic history of these Oligo-Miocene carbonates from the NCB remain poorly 102 constrained, mostly due to the scarcity of rock samples available from offshore exploration 103 wells (mainly cuttings). Detecting and quantifying the impact of depositional and diagenetic 104

105 processes on acoustic properties is of paramount importance for correctly interpreting the seismic expression of carbonate systems and reservoirs (Fournier and Borgomano, 2007; 106 Teillet et al., 2020). In the present work, we therefore define a well-constrained depositional 107 108 and diagenetic framework from nearby, age-equivalent outcrops, in order to assess the combined impact of carbonate production (heterozoan vs. photozoan) on the diagenetic 109 evolution and acoustic properties of carbonate reservoirs (Fig. 2). The main goal of this study 110 is thus to emphasize the interplay and feedbacks existing between primary sedimentary 111 attributes, diagenetic paths and the resulting petroacoustic signature of distinct carbonates 112 113 types. To achieve this objective, the integrated approach implemented in this work (sedimentology, diagenesis, laboratory petroacoustics and numerical modelling) aims at (i) 114 constraining the sedimentological and diagenetic framework of the Oligo-Miocene 115 116 carbonates sequences from the NCB, (ii) identifying the key parameters controlling the evolution of pore volume, pore network architecture and mineralogy in both heterozoan and 117 photozoan carbonates, (iii) quantifying the relative impact of depositional features (texture, 118 initial mineralogy) and diagenetic transformations on acoustic properties of reservoirs related 119 to both carbonate associations. 120

121 **<u>2. Geological setting</u>**

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123 This work primarily concerns carbonate outcrops of Oligo-Miocene age from the 124 Northwest Shelf of Australia. The analysis focused on outcrops from the Cape Range 125 peninsula, located in the onshore part of the Exmouth Sub-Basin (NCB, Western Australia, 126 Fig. 1).

127 The NCB is the largest hydrocarbon province of Australia. Since the onset of hydrocarbon 128 exploration in 1953, 145 GL of oil, 52 GL of condensate and 207 BCM of gas have been 129 produced, mostly from Mesozoic age reservoirs. The sedimentary cover of the NCB, 130 overlying the Jurassic and Cretaceous hydrocarbon-bearing reservoirs, is mainly composed of heterozoan ramp to photozoan carbonate platform deposits from Late Cretaceous to 131 Quaternary ages (e.g. Hocking, 1987; Cathro et al., 2003). Onshore exposures of the Oligo-132 133 Miocene carbonate sequences of the NCB occur across the 20-30 km wide and 60 km long Cape Range peninsula, essentially within canyons and quarries (Fig. 1B). The Cape Range 134 peninsula is located on the margin of the Exmouth sub-basin which is part of the NCB 135 (Hocking, 1987). This basin evolved from a rifted margin (late Triassic to Early Cretaceous) 136 to a passive margin (Early Cretaceous to present), and has been associated with a moderate 137 138 and fairly constant subsidence during the Cenozoic (Cathro et al., 2006; Czarnota et al., 2013). The formation of the Cape Range peninsula topographic high resulted from the 139 140 Exmouth sub-basin inversion which commenced during the Miocene. This tectonic inversion 141 is characterized by the formation of several *en-échelon* SW-NE oriented open folds, such as the adjacent Rough Range and Giralia Anticlines, resulting from strike slip reactivation of en-142 échelon basement faults (e.g. Hocking et al., 1987; Fig. 1B). 143

144 The Oligo-Miocene carbonate system of the NCB has been previously interpreted as a nontropical (temperate to sub-tropical carbonates) low-gradient, distally-steepened ramp 145 prograding to the north-west. These results were obtained from the analysis of offshore wells 146 and seismic (Young et al., 2001; Cathro et al., 2003; Moss et al., 2004) as well as from 147 sedimentological studies of exposed sections in Cape Range (e.g. Condon, 1955; Collins et 148 149 al., 2006). Regionally, the Oligo-Miocene carbonate have been subdivided into three distinct formations (Figs. 1B and 3): the Late Oligocene to Early Miocene Mandu Limestone and 150 Tulki Limestone and the Early to Middle Miocene Trealla Limestone (Fig. 3A) (Condon et 151 152 al., 1955; Quilty, 1980; Van der Graaff et al., 1980; Apthorpe, 1988; Riera et al., 2019) that, in the onshore part of the system, rest unconformably on the Tulki (Fig. 1C). The age of each 153 of these formations has been established from foraminifer-based bio-stratigraphy 154

155 (Chaproniere, 1975, 1984; Quilty, 1977; Apthrope 1988; Riera et al., 2019). The Mandu Limestone was deposited from the Late Oligocene to Early Miocene, including foraminifer 156 zones P19-21 to N6 (Blow, 1969 nomenclature), which implies an age between 26 and 18 157 158 Ma. The age of the Tulki Limestone range from zones N6/7 and – possibly – up to the N8 zone on the eastern flank of Cape Range (Chaproniere, 1984) (Chaproniere, 1984), 159 corresponding to an age from around 18 to 15 Myrs. Finally, the age of the basal Trealla 160 161 Limestone ranges from upper Burdigalian in the southern Cape Range peninsula to the upper Langhian to the North (Riera et al., 2019). The NCB is characterized by one of the most 162 163 extensive open-access offshore 2D/3D high-resolution seismic data and well data coverage. Such offshore datasets provide excellent insight into the stratigraphic architecture of the 164 Oligo-Miocene carbonates (Cathro et al., 2001; 2003; Moss et al., 2004; Ryan et al., 2009; 165 166 Rosleff-Soerensen et al., 2012; 2016; Smith, 2014; Saqab and Bourget, 2016), but only cover the more distal deposits (mid to outer ramp, slope and basin). The proximal parts of the basin 167 are poorly penetrated by exploration wells and has almost no seismic coverage. However, 168 169 onshore outcrops, along the anticline folds uplifted and exposed during the Middle-to-Late Miocene inversion of the margin (Hocking et al., 1987; Cathro et al., 2004) allow to study the 170 most proximal parts of the system. Unfortunately, onshore studies of the Oligo-Miocene 171 sequences of the NCB are few, so that the current knowledge is limited to their broad age and 172 a generalized facies and stratigraphic evolution model (Collins et al., 2006). 173

In the studied outcrops, the Mandu Limestone (up to 300m thick; Hocking et al., 1987) exhibits an upward vertical change in facies from white to yellowish, fine grained marly and fine calcarenite (Lower Mandu) to fine-grained limestone with interbeds of larger foraminiferal packstone to floatstone (Upper Mandu) (Fig. 3A-C). The Tulki Limestone (up to 129m thick) consists of a hard, thick-bedded limestone interpreted as a proximal, proximal equivalent of the Mandu Limestone (Chaproniere, 1984, Fig. 3D). The Trealla Limestone is a 180 diachronous carbonate formation (Riera et al., 2019) which overlies the Tulki Limestone unconformably (Quilty, 1977; Chaproniere, 1984; Hocking et al., 1987; Fig. 1C). Locally, 181 previous biostratigraphy dating suggests a time gap between Tulki and Trealla on the order of 182 around 1 Myr, at least in the Giralia anticline (Chaproniere 1984). The Trealla Limestone is a 183 hard, white to cream crystalline limestone (Fig. 3F-G). A significant uplift (200-300m) and 184 subsequent emersion of the Cape Range area occurred during the Late Miocene, as a result of 185 deeper fault reactivation (Hocking et al., 1987). Outcropping thicknesses for the Mandu Tulki 186 and Trealla formations in the Cape Range region average 80 m, 30 m and 20 m, respectively 187 188 (Fig. 3A) (Chaproniere, 1984; Collins et al. 2006).

189 **<u>3. Database and Methods</u>**

This study is based on a multiscale (microscale to kilometre) and multidisciplinary 190 approach, (Fig. 2), including field sedimentology, petrographic analysis of thin-section 191 192 (microfacies, composition of the biota, diagenetic features), cathodoluminescence, carbon and oxygen stable isotope measurements, calcimetry, laboratory P and S-wave velocity 193 measurements, helium porosity measurements and micro-CT scan imaging. The field-work 194 focused on four main outcrops, located on the Cape Range peninsula (Fig. 1), Shothole 195 Canyon (22.06319S;114.01199E, "1" in Fig. 1B), Charles Knife Road (22° 6'57.87"S;114° 196 2'20.10"E, "2" in Fig. 1B), Unnamed Quarry 1, (21°59'36.16"S;114° 5'37.98"E, "3" in Fig. 197 1B) Unnamed Quarry 2 (21°59'50.20"S;114° 04'54.50"E, "4" in Fig. 1B). 198

- 199
- 200 **3.1 Carbonate petrography**

During field-work, 47 rock samples were collected from the 4 sites studied, including sites previously studied by Collins et al., (2006) as well as the previously unstudied Unnamed Quarry 2. The Mandu and Tulki limestones facies were described mainly from Shothole Canyon and Charles Knife Road, while expose facies descriptions of the Trealla formation were derived from all main localities. Trealla age-equivalent deposits also outcrop in Unnamed Quarry 2 (Fig. 1B), not previously described. Thin sections have been prepared from all 47 samples and subsequently inspected using optical microscope under polarized light as well as under cathodoluminescence in order to characterize both depositional and diagenetic features (Fig. 2).

High resolution scanning of each thin-section was carried out at UWA-CMCA using a Leica 210 (Aperio) Scanscope XT that provided ×20 magnification digital slides. From these scans 211 detailed point counting analyses has been performed using JMicroVision (Roduit, 2007) 212 213 software in order to define the dominant pore type (DPT), as well as the relative proportion of macroporosity, the quartz content, and the micrite/sparite ratio of each sample. At least 400 214 points selected randomly on the image were required for each scan to obtain statistical 215 216 relevance/mean stationarity of the considered classes (sparite, micrite, allochem, pore type, 217 quartz). Micropores have been defined as pores with diameter smaller than 10 µm (Cantrell and Hagerty, 1999) while macropores have diameter higher than 10 µm. Vugs are defined as 218 dissolution voids which cross-cut original grain boundaries and usually larger than the mean 219 grain size. 220

221

222 3.2 Laboratory measurements

223 Physical properties

Porosity and velocity measurements were performed on plugs at the Department of Earth Sciences of the Vrije Universiteit Amsterdam. The plugs (diameter: 3.8 cm; length: 6 cm) have been drilled on oriented samples, the same used for petrography, along their vertical direction. Porosity measurements have been performed using a helium pycnometer (AccuPyc 1330, Micrometritics). The pore volume is calculated from the ratio of the plug's bulk density on the grain density (following Kenter & Ivanov, 1995 methodology). Grain density is estimated from the volume of helium filling the pore space and bulk density is estimated fromthe volume and weight of the plug.

Among the 47 sample, only 39 allowed to obtain consistent P and S-waves velocity 232 233 measurements. Velocities were measured using the Verde Geosciences High Pressure Measurement System (HPMS) under dry conditions with increasing isotropic confining stress 234 of 2.5, 5; 10; 20 and 40 MPa, where most velocity values reach a plateau (see Fig. S1). 235 During measurements plugs are placed in an impermeable rubber jacket and immersed in a 236 confining oil which pressure is gradually increased. Two piezoelectric transducers 237 238 (emitter/receiver) generating ultrasonic source of 1 MHz frequency were used for the measurements performed along the plug axis. Measurements are based on the picking of 239 wave first-break arrival, using a 3% of maximum amplitude threshold. 40 MPa is a classical 240 241 value for confining pressure used in a majority of petrophysical studies. This value is greater 242 than the highest burial stress experienced by the studied rocks, and has the advantage to make comparison of our data with other studies more valid. 243

Three-dimensional micro CT-scan acquisition (100 µm resolution and 0.1 mm section spacing: along the plug sample) have been performed in TOTAL CSTJF in Pau, in order to reassess the dominant pore-type of each sample with 3D data, detect the occurrence of possible plug scale heterogeneities and the presence of open discontinuities (fractures, stylolites).

249

250 Mineral geochemistry

CaCO₃ content has been quantified by using an automated pressure calcimeter in the CEREGE laboratory (Aix-Marseille University) from rock-powder samples, drilled directly from the petrophysics plug samples. Measurements of stable isotopes ratio for Carbon and Oxygen (δ^{13} C and δ^{18} O) have been carried out on 42 powder samples acquired from 255 millimeter-size drill bit and using an inductively coupled plasma mass-spectrometre (ICP-MS), at the Erlangen-Nürnberg University. Stable isotope data provide constraints in the 256 interpretation of diagenetic environments and the paragenesis of the carbonates investigated. 257 258

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3.3 Petroacoustic modelling: the EPAR approach

Effective property models, such as differential effective medium (Norris, 1985; 261 Zimmerman, 1991), have been shown to be a powerful tool in modelling velocity-porosity 262 transforms and elastic moduli-velocity transforms, relating them to pore structure (Xu and 263 Payne, 2009; Fournier et al., 2011; 2014; Cilli and Chapman, 2020; Singh et al., 2020). The 264 Equivalent Pore Aspect Ratio (EPAR) was defined by Fournier et al. (2011) as the aspect 265 ratio of pores from an equivalent elastic medium consisting of a non-porous calcitic host with 266 spheroidal inclusions, which displays the same elastic moduli as the one measured on the 267 actual rock. It is based on DEM (Differential Effective Medium) models and allows the 268 combined effect of porosity, mineralogy and pore-type onto acoustic properties to be 269 separated and quantified. The EPAR derived from the bulk modulus is named $\alpha_{\rm K}$ (Fig. 2). 270 The bulk modulus of the mineral phase, used in DEM computations, has been calculated by 271 using the mean of the Hashin and Shtrikman (1963) bounds and by considering the calcite 272 and quartz fractions estimated from CaCO₃ content (calcimetry) and point counting. The α_K 273 value of each sample is determined by matching the measured bulk moduli and porosity with 274 the DEM models (Fig. 2 and Figure S1). Consequently, each of the 41 experimental couple of 275 values (K; Φ) falls on one of these model curves, indicating its α_{K} . It is worthy to mention 276 277 that the α_K value is not a direct estimate of the actual pore aspect ratio (which can be heterogeneous), but represents a proxy of the pore network stiffness (Fournier et al., 2018), if 278 all the pores were perfectly similar in shape. 279

280 Considering a source with a main frequency of 1MHz and the values of measured P-wave 281 velocities, wavelengths should range from 3 to 6 mm. A few samples exhibit macropores 282 which are of >5-10mm scale, which consequently don't fall into the effective medium 283 domain regarding our measurements (these samples have been highlighted in Table 2).

284

285 **<u>4. Results</u>**

4.1. Depositional facies and paleoenvironments

287 4.1.1 Mandu Limestone formation (Facies association F1)

The Mandu Limestone is mainly composed of micritic wackestone to packstone facies 288 (Facies association F1; Fig. 3B and Fig. 4A-B; Table 1A). In F1 grains are mainly Large 289 Benthic Foraminifers (LBF, lower arrow in Fig. 4A), Planktonic Foraminifers (globigerinids), 290 and scarce coralline algae and F1 can show a variable amount of quartz grain (upper arrow in 291 Fig. 4A; up to 14% based on calcimetry Fig. 5A). Mandu Limestone can be separated in two 292 intervals (Lower and Upper Mandu Fm., Facies F1A and F1B; Fig. 3A-3B), based on 293 texture, micrite proportion and the faunal content. Foraminiferal assemblage and other 294 295 allochems indicates a depositional environment changing from the lower oligophotic zone (F1A) to mesophotic zone (F1B) (See Appendix A and Figure S2 for a more detailed facies 296 description and depositional environment interpretations). 297

298 4.1.2 Tulki Limestone formation (Facies association F2)

The transition between the Mandu and Tulki Limestones successions is gradual and shows a decrease in matrix micrite content, terminating a long-term shift from > 80% in the lower Mandu to < 10% at the base of the Tulki (Fig. 4C and Fig. 5B, Table 1A). The Tulki Limestone is characterized by coarser bioclastic packstone to grainstone, locally crudely 303 cross-bedded (Facies association F2, Table 1A), with the grains dominated by red algae, 304 echinoderms, bryozoans, rodoliths (black arrow in Fig. 4C), reworked LBF (see Fig. S2) and 305 shallow benthic foraminifers (see Appendix A). The foraminiferal assemblage and texture 306 indicates that the Tulki Limestone accumulated in a comparatively shallow, high-energy 307 environment, well within the euphotic zone.

308

4.1.3 Trealla Limestone formation (Facies association F3)

The Trealla Limestone records a marked change in terms of texture and grain 309 composition as compared to the Mandu-Tulki succession, with high vertical and lateral 310 variability. Two broadly defined facies have been recognized within Facies association 3 311 312 (F3): (i) Scleractinian-Halimeda floatstone with a red algal-foraminiferal (mainly represented by large miliolids) packstone matrix (Fig. 4D) and (ii) red and green algal-foraminiferal 313 packstones to grainstones (Fig. 4E). The faunal and floral content (see Appendix A) of both 314 315 facies indicates that Trealla Fm. limestones accumulated in a very shallow environment (< 10m), within the euphotic zone. 316

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- 318 **4.2 Diagenetic features and pore types**
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In carbonates rocks especially, pore volume and pore types are not only correlated to initial sedimentary processes but are strongly linked to the diagenetic transformations they have undergone from deposition to exhumation. The rocks studied in this work present a range of pore types including intercrystalline micropores (within matrix and allochems) to molds, intergranular, and vuggy macropores.

325 4.2.1 Mandu Limestone formation

326 Due to the dominantly mud-supported depositional texture of facies association F1, 327 the Mandu Limestone show very scarce sparite cements, except in a few intraskeletal pores 328 (Fig. 6A) and within mode-1 micro-fractures (SC) (Fig. 7A). Mandu is characterized by compaction markers such as stylolites/micro-stylolites, particularly in the micritic matrix 329 (Fig. 6A red arrows) and compaction mode-1 micro-fractures affecting essentially the LBF 330 331 (Fig. 6A black arrow). Micritisation and partial grain dissolution affects LBF skeleton, particularly the Lepydocyclinids (Fig. 6B red arrow). Cathodoluminescence observations 332 showed that the micrite matrix and micritized allochems are luminescent (dull-orange 333 luminescence), and revealed an additional phase of dark blue luminescent to non-334 luminescent, drusy to mosaic sparite cement (SC, Fig. 7B). This particular sparite cement is 335 336 located mainly in the recrystallized micrite area (Fig. 7B) and in the mode-1 fractures. Porosity in the Mandu Limestone is dominated by microporosity in micrite matrix and in 337 micritized grains (mostly peloids and lepidocyclinids). Residual macroporosity in 338 339 intraskeletal pores and mode-1 microfractures after partial occlusion by calcite cement is 340 uncommon (Figs. 7A and 6B).

341

1 **4.2.2** Tulki Limestone formation

342 In the Tulki Limestone, pervasive compaction features such as large stylolites and concavo-convex grain contacts are observed (arrows in Figs. 6C and 6F). As a result of such 343 a compaction, few large blocky sparite cements have been able to develop in the reduced 344 intergranular pore space. In most of the grainstone samples, the remaining intergranular pore 345 space is entirely filled by two distinct phase of sparry calcite cement: (i) perigranular, 346 isopachous fringes of prismatic sparry calcite cement (ISC) (Figs. 6D, 6E and 7C) displaying 347 a dull-orange luminescence (Fig. 7D), and (ii) dark-blue luminescent to non-luminescent, 348 drusy to mosaic, blocky calcite cement (SC) in intergranular position and inside the fractures 349 (Figs. 7C and 7D), similar to those observed in the Mandu Limestone. Locally few 350 351 anisopachous micritic and fibrous perigranular cements occur (arrows in Fig. 7D), suggesting an early diagenetic phase happening in a vadose and/or marine environment (e.g. Flügel,2010).

In the Tulki Limestone, pore types are characterized by a dominance of various levels of 354 355 intragranular microporosity within micritized grains which can lead to moldic/vuggy porosity (Figs. 6D and see 8A). In a few samples, where compaction features are less pronounced, 356 intergranular porosity may dominate. Dissolution features affecting grains (vugs and moldic 357 pores after bioclast leaching) or locally even matrix have been observed, particularly in the 358 upper part of the Tulki Limestone (Figs. 8A and 8B). Such dissolution pores (molds and 359 360 vugs) are commonly filled by the non-luminescent blocky calcite cement (SC). Like in the Mandu Limestone, peloids and micritized bioclasts exhibit a dull-orange luminescence (Fig. 361 7D). In grainstones, especially where the perigranular fringe of prismatic calcite is lacking, 362 363 concavo-convex contacts between grains are often observed (Fig. 6C). Additionally, stylolites 364 are common, particularly in packstone intervals (Fig. 6F). No crosscutting relationship between SC and stylolites/concavo-convex grain contact has been observed, suggesting that 365 SC calcite formed after development of the burial features. 366

At the macroscopic scale, another diagenetic feature is a rock pervasive oxidation resulting in 367 a reddish colour (Figs. 3D and 3E). It is important to note that this oxidation does not affect 368 outcrops of the overlying Trealla rocks, at least in studied areas. From thin sections we 369 370 established that iron oxide take place preferentially and mostly inside the intraskeletal spaces 371 (Figs. 6C and 6E). It is further found as fringes lining the boundary between perigranular prismatic cement (ISC) and the later non-luminescent blocky sparite cements (SC) (arrow in 372 Fig. 6E). This suggests that iron oxide deposition took place relatively early in the diagenetic 373 374 history, prior to the blocky sparite cementation and shows it is not related to a modern, telogenetic meteoric process. Iron oxides are also found lining the walls of the moldic pores 375 and dissolution vugs (Fig. 8B), thus suggesting that they have circulated during or after this 376

377 dissolution phase. Since the dissolution vugs have been filled by both ISC and SC, the dissolution phase is therefore interpreted to have occurred before sparry calcite cementation., 378 thus at least prior to the final exhumation of these rocks (see section 4.7). 379

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- 381

4.2.3 Trealla Limestone formation

The Trealla Limestone, contrary to Mandu and Tulki, is characterized by a 382 widespread preservation of the primary intergranular porosity, especially in the grainstone 383 facies. All samples show little-to-no intergranular cementation and a dominance of point-384 contact between grain (Fig. 6G and see Fig. 13F), with no pressure-solution features. To a 385 larger extent, no indication of mesogenetic diagenesis (sensu Choquette and Pray, 1970) has 386 been observed. Grains are generally rimmed by a thin isopachous fringe of equigranular 387 sparry calcite cement (Fig. 6G red arrow), consistent with a formation within a meteoric 388 phreatic lens (e.g. Tucker and Wright, 2009). Small areas of perigranular fibrous cement have 389 390 also been observed. Aragonitic bioclasts (corals and green algae) are commonly leached, forming large moldic to vuggy pores lined by a similar fringe of equigranular calcite or -391 rarely – completely filled by mosaic sparite (Fig. 6H). 392

Under cathodoluminescence the equigranular calcite is non-luminescent in its inner, early 393 part and exhibits a dull-orange luminescence in its outer boundary ("ZSC" on Figs. 7E and 394 7F). Such zoned sparry calcite cements (ZSC) should be therefore distinguished from the 395 isopachous sparry calcite (ISC) observed in the Mandu and Tulki formations. Small areas of 396 dark-blue luminescent to non-luminescent sparry cement developed in the intergranular pore 397 398 spaces (Fig. 7F), is plausibly equivalent to the SC non-luminescent cement observed in the Tulki Limestone. As a result of limited cementation and lack of significant compaction 399 features, some samples still preserved a large amount of initial intergranular porosity (see 400 401 Fig. 13F).

402

403 **4.2.4 Outcrop-scale structures and karstification analysis**

The "Unnamed Quarry 2" (Fig. 1B) exposes a decametre scale outcrop displaying tectonic 404 structures and karstic features at the top of the Tulki limestones (Fig. 8C). Although, it is 405 difficult to precisely (bed to bed) correlate the stratigraphy, this quarry exhibits an uneven 406 407 contact between LBF rich yellowish Tulki facies (F2) and white Trealla coral rich facies (F3) ("1" and "2" in Fig. 8C). The outcrop is affected by a N-S striking sub-vertical fault with a 408 normal component (in yellow Fig. 8C). This fault shows vertical karst features (macro-scale 409 dissolution) along its fractures in the associated damage-zone. A stratiform bed (of variable 410 thickness) of pseudo-breccia embedded in a red "terra rossa" matrix is found (Fig. 8D) at the 411 interface between the Tulki and Trealla formations, suggesting that a subaerial exposure and 412 karstification phase occurred prior to Trealla deposition. Immediately below the karst-breccia 413 bed, the limestone is affected by decimetre-to-meter scale dissolution voids, mostly 414 415 horizontal and filled with calcite cement (Fig. 8E). Such diagenetic features as well as the pseudo-breccia bed are both cross-cut by the fault (and associated vertical karst) that 416 postdates it, thus supporting the assumption that these karstic pseudo-breccias are not related 417 to modern telogenetic processes, in contrast to the quaternary pisolitic material infilling voids 418 within the Trealla, observed by Hocking et al. (1987). 419

420

421 **4.3 Carbon and oxygen stable isotopes ratios**

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The stable isotope signature of bulk-rock carbonates has been used to interpret the origin of the main diagenetic transformations (Fig. 9A). For all measured samples, the isotopic signatures are negative and depleted in both ¹⁸O and ¹³C as compared to the Miocene marine calcite values (e.g. Veizer et al., 1999; Fig. 9A). Isotope ratios exhibit a narrow δ^{18} O

range, averaging around -6 % VPDB, in combination with more scattered δ^{13} C values, 427 ranging from -1 to -9 % VPDB. This suggests that these rocks: (i) have undergone diagenetic 428 transformations, resulting in a loss of their initial marine geochemical signature and (ii) that 429 430 they underwent diagenetic transformations (neomorphism and/or cementation) in a meteoric environment influence, as suggested by low negative δ^{13} C (e.g. Hudson, 1977; Allan and 431 Matthews 1982). More particularly, the two measurements of the ZSC sparry calcite cements 432 from the Trealla limestones revealing highly negative $\delta 13C$ values, ranging from -7.07 to -433 8.67%, while the corresponding δ 18O values range from -5.53 to -6.05%, thus strongly 434 435 suggesting a meteoric origin for ZSC cements. One measurement made specifically from oxidized grain from the Tulki also display strongly negative value, suggesting that oxidation 436 and its associated dissolution phase, also took place under strong meteoric influence. 437

Noteworthy, δ^{13} C values from the Tulki and Mandu limestones shows decreasing upward 438 values base to top (Fig.9B). This strongly suggests that such a meteoric δ^{13} C signature is not 439 related to modern meteoric alteration, which would have uniformly all outcrops investigated, 440 441 but rather to earlier processes occurring prior to the final exhumation of these rocks. Furthermore, an abrupt negative shift in δ^{13} C values occurs at the top of the Tulki Limestone 442 (Fig. 9B), with δ^{13} C values down to -8 %, which is consistent with the occurrence of a 443 subaerial exposure at the top of Tulki formation (e.g. Quilty, 1977; Chaproniere 1984 and 444 Condon 1968; see also 5.1). Finally, low δ^{13} C (-6 to -7 %) in the Trealla Limestone, are close 445 446 to those of (ZSC) cements and fits well with its very shallow water deposition environment and are indicative of a later phase of meteoric diagenesis. 447

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449 **4.4 Pore type classification**

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451 A pore type classification has been developed based on a semi-quantitative analysis of 452 thin-sections and CT-Scan images. In the studied samples, macropore space consists 453 dominantly of intergranular, moldic and vuggy pores (Table 1B). Primary skeletal 454 macropores are rare and do not represent a significant proportion of the macroporosity. Intergranular pore spaces are mostly partially filled by calcite cements. Most of the samples 455 456 display varying amounts of intercrystalline micropores located within micritized skeletal grains and/or within micritic matrix. The proportion of macropores has been semi-457 quantitatively estimated from point-counting on thin-sections high-resolution scans. 458 Microporosity has been estimated by calculating the difference between measured total 459 helium porosity and macroporosity determined by point counting. CT-Scan images of plugs 460 461 have been used to assess possible pore type heterogeneities at the plug scale as well as to detect open discontinuities such as stylolites, fractures or vugs that have not been captured by 462 thin-sections. Samples have been classified into 5 groups according to their dominant pore 463 464 type (Table 1B): (i) DPT1: dominant intercrystalline microporosity with scarce or lacking macropores, (ii) DPT2: mixed microporosity and intergranular macroporosity, (iii) DPT3: 465 dominant intergranular macroporosity, (iv) DPT4: dominant moldic and/or vuggy 466 467 macroporosity. Samples with occurrences of open stylolites or micro-fractures, have been classified as above, with the addition of the suffix S (DPT1S, DPT2S and DPT3S). 468

469

470 **4.5 Mineralogy, porosity and ultrasonic velocities**

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Measurements of CaCO₃ content (calcimetry) performed on 20 samples from the Mandu and Tulki Formations (Fig. 5A) revealed that a significant amount of non-calcite material (up to 14%), may exist in the bulk rock, especially in the lower Mandu samples. This proportion rapidly decreases and become negligible at the transition with the Tulki Limestone. Considering the percentage of quartz estimated from point counting (Table 2), the non-calcitic fraction is assumed to be mainly composed of quartz grains. 478 Vp values at an effective stress of 40 MPa are highly scattered, ranging from 2351 m/s from to 6402 m/s. These velocities are mainly consistent, though slightly higher, than the wireline 479 log data (performed at dm to m scale) from wells penetrating the age-equivalents limestones 480 481 in the offshore part of the NCB (~ 2000-5000 m/s; Wallace et al., 2003; Shragge et al., 2019). These offshore equivalents (Cathro et al., 2003; Wallace et al., 2003; Moss et al., 2004) 482 largely corresponds to the finer-grained facies F1A. Porosities values are also highly 483 scattered between 0 and 35% (Table 2), with average porosities of 18% (Mandu Fm; F1), 484 2.8% (Tulki Fm; F2), and 8.3% (Trealla Fm; F3). The highest porosity values are found in the 485 486 microporous samples from the Mandu Limestone (e.g. Msh-A15 27.5%). Ultrasonic velocities display a nearly linear decreasing trend with increasing porosity but with a strong 487 scattering of values at a given porosity, averaging 1000 m/s for P-waves (Fig. 10A) and 500 488 489 m/s for S-waves (Fig. 10B). This variation in velocity values at a given porosity is not fully 490 explained by the depositional texture or facies. Tight (porosity <1%) samples exhibit highest P-wave velocities, ranging from 6000 to 6500 m/s, which is consistent with values measured 491 492 in pure calcite.

Dynamic bulk and shear moduli have been computed from Vp, Vs, and bulk density 493 measurements (Table 2,). In tight samples, both moduli values range are within the 494 magnitudes known for pure calcite (K_{calc}=70 GPa, µ_{calc}=30 GPa, e.g. Mavko et al., 2008). 495 496 They are found to be negatively correlated with porosity, but again with significant scattering for a given porosity value (Fig. 11). Depositional parameters along with pore types seem to 497 control bulk moduli-porosity relationships. Indeed, the higher K values of the Trealla 498 limestones, as compared to the Mandu and Tulki, is clearly correlated with dominant pore 499 500 types, with a moldic to vuggy pore regime (DTP4) dominant in the Trealla and matrix and intragranular microporosity dominant in the Mandu and Tulki limestones (DTP1 to DTP2) 501 (Fig. 11A and 11B). 502

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504

4.6 EPAR modelling and correlation with dominant pore type

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In order to explain the cause of the observed scattering in velocity and bulk modulus at a given porosity and try to predict accurately pore type information from indirect measurements, we applied the EPAR approach, by implementing the so-called α_{K} proxy (Fournier et al. 2018). In the limestones studied, the obtained equivalent pore aspect ratio (α_{K}) range from 0.03 to 0.53 (Fig. 12). It is interesting to point out that the α_{K} -porosity plot shows that no correlation exists between α_{K} and porosity, thus suggesting this parameter is not controlled directly by pore volume but more likely by pore network structure.

In contrast, dominant pore types, determined from quantitative point counting (Table 2; colours in 513 Figs. 11B and 12), are fairly well discriminated by the $\alpha_{\rm K}$ proxy, with: (i) samples characterized 514 by a dominant intercrystalline microporosity (DPT1) showing α_K values between 0.03 and ≤ 0.2 , 515 for a wide range of porosity value from 0.5 to 35%, (ii) samples characterized by dominant 516 517 moldic to vuggy pores all display values >0.2 (DTP4); (iii) samples characterized by dominant intergranular pores show α_K values mainly around 0.3; (iv) samples with mixed 518 519 micro/intergranular porosity type show α_K values mainly around 0.2 (DTP2). (v) samples affected by open discontinuities (DPT1S and DPT3S) exhibit α_K values lower than their non-fractured 520 521 counterpart, regardless of the initial DPT.

It appears, therefore, that the α_{K} variation is primarily related to the dominant pore type in each sample, i.e. related to depositional and diagenetic processes. However, it is important to note that for low porosity values (less than 5%) the dominant pore type show little correlation with the EPAR variation (grey zone in Fig. 12). This is explained by the fact that, DEM models tend to converge at low porosity (see Fig. S3) thus leading to increased uncertainty in α_{K} computation and therefore inefficient discrimination of pore structures.

528

529 **4.7 Paragenetic sequence influenced by subaerial exposure/erosion phase**

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Based on outcrop data, petrographic analysis and stable isotope results, a paragenetic 531 532 evolution of the Oligo-Miocene carbonates from Cape Range is proposed (Fig. 13A-D). This paragenetic sequence and preceding sedimentological results reveals the major differences 533 between the Mandu-Tulki and Trealla limestones in terms of both depositional features and 534 diagenetic evolution. In particular, compaction and burial features, are prominent in the 535 Mandu and Tulki formations, whereas they are completely lacking within the Trealla 536 formation (Fig. 13E-F). In spite of their similar dominant depositional texture (packstone-537 grainstone), distinct patterns of diagenetic evolution led to dramatically different porosity, 538 pore network and acoustic properties. 539

Due to their limited total thickness in the Cape Range (≤ 100 m), the Tulki and Trealla would 540 be expected to have experienced comparable burial depth and thus display similar diagenetic 541 processes and compaction features (Fig. 13A). Compaction features, such as stylolites observed 542 in the Tulki (and Mandu) Limestone formed during burial, under lithostatic stress. Although no 543 544 consensus exists regarding the minimum depth for their development, laboratory and in-situ 545 studies suggest that stylolites do not require very high burial depth to form. Their nucleation has 546 been estimated to occur at depths around 100m (Koehn et al., 2012) whereas actual stylolite planes have been shown to develop at depths ranging from 200 to 1000m (e.g. Nicolaides and 547 Wallace; 1997; Kroon 2017; Bruna et al., 2018). The frequent and consistent occurrence of 548 stylolites and concavo-convex grain contacts throughout the Tulki limestones, compared to 549 the lack of any significant compaction features within the Trealla limestone, therefore suggest 550 that an emersion occurred at top of the Tulki and led to the erosion of at least several 10s of 551 meters of sediments, prior to deposition of the Trealla Limestone (Fig. 13A). Such an event 552 has been suspected by Condon et al. (1955), Quilty (1980) and Chaproniere (1984), who noted 553 thickness variations of the Tulki Limestone along Cape Range, the presence of an iron-rich layer 554

555 associated at an erosional surface at the Top-Tulki and, furthermore, mentioned a possible slight angular unconformity between the Tulki and Trealla formations. From borehole date, they also 556 noted an "intense alteration" of the Top-Tulki in onshore and proximal-offshore wells Muiron-1 557 558 and Ningaloo-1. Such an erosion, would have led to the erasure of the less-compacted part of the Tulki. The subsequent uplift of the Cape Range region during the Late Miocene did not allow 559 for the deposition of thick sedimentary successions, thus explaining the scarcity of 560 compaction features within the Trealla Limestone (even at its base). The occurrence of a 561 562 major subaerial exposure event affecting the top of Tulki Limestones as proposed here is supported by: 563

- (i) Karstic features affecting the top of Tulki limestones in the « Unnamed Quarry 2 »
(Fig. 8) which are sealed by the Trealla Limestone.

- (ii) The occurrence of iron oxide deposits postdating both the perigranular (ISC) cement, and
the dissolutions voids, (but predating both blocky SC cement, Figs. 6D and 6E) and affecting the
limestones below the top Tulki strongly supports that this oxidation takes place during this event.
It also explains why there is no such oxidation, in the Trealla Limestone, even in the currently
open pore network.

571 - (iii) The highly negative $\delta^{13}C$ shift between the at the Tukli-Trealla boundary.

572

573 The scenario outlined above implies a short-lived fall in relative sea level around the onset of the Middle Miocene. This relative drop may have been driven by an early stage of the North-574 Australian margin compression, which started with the collision with Timor during the 575 Middle Miocene (e.g. Cathro, 2003; Jitmahantakul and McClay, 2013) with modest 576 transpression, even in the Cape Range area. Indeed, evidences of minor folding events during 577 the Middle Miocene have been found offshore, North of Cape Range, in the Browse and 578 Barcoo sub-basin (e.g. Rosleff-Soerensen et al. 2016, Wilson et al., unpublished data). 579 Onshore Cape Range, where possible episodic deformations have been documented since the 580

Mid-Miocene (e.g. Keep et al., 2012, Condon, 1968, Quilty 1980). At the scale of Cape Range, even a slight overall tilt of 1° of the bedding could have led to the uplift up to 300°m, on the other (eastern) side of the peninsula. The actual main phase of the tectonic inversion of the North West Shelf basins starts latter (10-7 Ma; e.g. Keep et al., 2007; Rosleff-Soerensen et al. 2016; Saqab et al., 2017). It is thus likely that the final uplift of the Cape Range region initiated soon after the deposition of the Trealla limestones, explaining their lack of compaction features.

The influence Middle Miocene glacio-eustasy associated with the rapid growth of the East 588 589 Antarctic Ice Sheet (Miller et al., 1991; Zachos & Bohaty, 2008; John et al., 2011), resulting in sea level fluctuations in the amplitude of 20-60 m (John et al., 2011, Miller et al., 2013) 590 cannot be excluded. Such eustatic variations could have resulted in repeated exposure of the 591 592 low angle ramp system along which the Tulki limestone was deposited. However, the limited 593 geographic extent of the subaerial exposure makes this explanation less probable. In both cases, significant erosion of the Tulki Limestone may have occurred during subaerial 594 exposition and/or during the subsequent Trealla marine transgression. 595

596

597 **5. Discussion**

598 **5.1 Impact of depositional and diagenetic processes on pore type evolution.**

599 5.1.1 Microporosity development

The present work revealed that Mandu/Tulki and Trealla Limestones in the Cape Range anticline are characterized by distinct carbonate grain producers as well as different path in diagenetic evolution. (i) Together, the Mandu and Tulki formations constitute the proximal part of an extensive heterozoan ramp system whereas the Trealla limestones forms a sub-tropical to tropical, photozoan system (Fig. 14). (ii) While the Mandu and Tulki formations are characterized by significant microporosity and by the frequent pressuresolution features (Fig. 13E), the Trealla limestones are dominated by intergranular macroporosity associated with moldic and vuggy pores and little-to-no compaction (Fig. 13F).

The most prominent diagenetic feature affecting the Mandu limestone is the development of 609 microporosity within carbonate matrix and grains. The significantly ¹³C-depleted signature of the 610 mud-supported carbonates from the Lower Mandu strongly suggests that microporosity 611 developed as a result of neomorphic transformations of micrite within a meteoric diagenetic 612 environment (Fig. 9A). Micrite neomorphism and associated microporosity in marine carbonates 613 614 have been commonly assigned to meteoric diagenesis occurring during prolonged phases of subaerial exposures (e.g. Leonide et al., 2014; Deville de Periere et al., 2017). Two distinct 615 phases of cementation (ISC and SC) affect the Mandu and Tulki limestones. The ISC cement was 616 most likely formed under a reducing marine or meteoric phreatic environment, whereas the SC 617 phase is more indicative of an oxidizing meteoric environment, as suggested by the ¹³C-depleted 618 signature of the SC-rich Tulki grainstones facies (Fig. 9A). Furthermore, the precipitation of iron 619 620 oxide postdating the ISC cements and but predating the SC infilling dissolution vugs (Fig. 6) below the karstified top of the Tulki unconformity allow the following inferences. (i) significant 621 622 dissolution processes occurred during this exposure and (ii) SC formed during the same phase, prior to, or at the beginning of, the Trealla limestone deposition, where it may explain the same 623 SC cement found in small amount in the Trealla intergranular pore space. As a consequence, the 624 lack of significant microporosity within the Trealla limestone and the ¹³C-depleted signature 625 626 of microporous mud-supported limestones from the Mandu Limestone strongly suggest that microporosity development mainly occurred during the top Tulki during subaerial exposure. 627 Additionally, grainstone samples from the Tulki Limestones, whose intergranular pores have 628 been occluded by SC cement, exhibit intragranular microporosity. This indicates that the SC 629 630 cementation phase post-dates the development of microporosity (e.g. Léonide et al., 2014).

Finally, it is interesting to note that in contrast to the onshore Trealla facies studied here, the main diagenetic processes involved in the pore amount and pore type acquisition in offshore ageequivalent Trealla are essentially controlled by compaction and burial cementation, (Wallace at al., 2003), largely similar to those described for the on-shore Mandu and Tulki formations described here. Indeed, these offshore deeper deposits are unlikely to have been exposed to meteoric processes, such as emersion-related diagenesis.

Although the impact of recent telogenetic processes on the acoustic signature of these rocks 637 cannot be entirely ruled out, the present data support that such an impact is likely to be 638 minimal. The preservation of a ddecreasing upward ¹³C values below the top-Tulki subaerial 639 exposure surface strongly suggests that no significant cementation and/or neomorphism occurred 640 in recent telogenetic environments. Such processes would have implied a geochemical signature 641 independent from the vertical position along the sedimentary succession. Moreover, no 642 observations in the field, and more importantly on thin section, have shown indication of modern 643 dissolution features. Only one significant dissolution phase has been observed (Fig. 8A-B) and 644 645 has been shown to have occurred prior to the final exhumation of these rocks (cf. 4.7). As a consequence, modern exposure seems at least to not have significantly modified the 646 microstructure of the limestones considered in this study. 647

648

5.1.2 Impact of the nature of carbonate production on the diagenetic evolution

649 A significant change in carbonate production occurred during the onset of the Trealla carbonate system, which have strongly impacted the dominant pore type acquisition. The 650 dominant Low-to-High magnesian calcite heterozoan carbonate factory (LBF, red algae, etc, 651 652 Fig. 14A) of the Mandu and Tulki limestones is replaced, in the Trealla limestone, by a photozoan carbonate production dominated by aragonitic biota (scleractinian and green algae, 653 Fig. 14B.) known to be a less stable mineral than calcite (e.g. Tucker 1990), hence more 654 prone to dissolution. Here, the shallow water environment of the Trealla limestone, 655 particularly exposed to early meteoric diagenesis (cf. 4.2), favored the development of mainly 656

657 moldic and vuggy porosity. In contrast, the foraminiferal and red algal limestones from the 658 Mandu and Tulki formations, developed only pervasive microporosity in both grains and 659 matrix, during subaerial exposition. Such a drastic change in carbonate factory between the 660 Mandu/Tulki and Trealla carbonate systems most likely reflect several factors; the most 661 important of which are listed below:

- (i) In the cause of the northward drifting of the Australian continent, the Cape Range
 moved to latitudes around 30° by the end of the Early Miocene (i.e. during Trealla
 deposition time; Fig. 1A), corresponding largely to the present day southernmost
 latitude of coral reef development in the South Pacific and Indian Oceans.
- (ii) Local or regional modifications in nutrient content (from mesotrophic to oligotrophic conditions) in sea-water (e.g. Pomar, 2001) and its physicochemical properties. The fact that possible evaporitic minerals (anhydrite, dolomite) have been found both onshore and in offshore wells in the Trealla time–equivalents deposits (Wallace et al., 2003; Gallagher et al., 2018) also points to that hypothesis.
- (iii) The subaerial exposure and erosion affecting the top-Tulki Limestone- including
 possible marine erosion during the Trealla transgression could have led to a
 flattened topography at the proximal margin of the Exmouth basin, favoring the
 development of an extensive, euphotic, platform or-low-angle ramp carbonate
 system (Fig. 14B).

677

5.2 Impact of pore network and pore shape on the acoustic signature: physical and geological interpretation of the α_K proxy

680 5.2.1 $\alpha_{\rm K}$ proxy variation in the light of pore stiffness

The α_K values can be explained by the soft or stiff acoustic behaviour of each DPT 681 class (Fig. 12). Indeed, EPAR modelling is based on the fact that different pore aspect ratios 682 have different stiffness which can be separated in two domains, "soft" pores or "stiff" pores. 683 The boundary between stiff and soft pore domain usually corresponds to so-called reference-684 pores whose aspect ratios range from 0.13 to 0.25, depending on the type and scale of 685 acoustic data used (Xu and Payne, 2009; Zhao et al., 2013; Hairabian et al., 2014; Fournier et 686 al., 2018). In the Cape Range database, the threshold value between these two domains would 687 be around 0.2, i.e. at the transition between dominant microporosity (DPT1) and dominant 688 moldic/vuggy pores (DPT4). Interestingly, this value is very similar to the one obtained in 689 slope and basinal tropical carbonates (Hairabian et al. 2014), suggesting that this threshold 690 691 might have actual physical significance for the pore network geometry/connectivity. The high α_K values (>0.2) are associated with samples whose pore space is dominated by moldic 692 and/or vuggy macropore (DPT 4), which match with the stiff pores domain (Fig. 12). This 693 result is consistent with the low pore compressibility expected for such type of roundish and 694 695 equant-shaped molds and vugs (Figs. 4 and 6). The stiff nature of such pore types has been previously documented in a variety of published carbonate case studies (e.g. Anselmetti and 696 697 Eberli, 1999; Saleh and Castagna, 2004; Fournier et al., 2014; Hairabian et al., 2014).

The low α_K values (<0.2) are associated with samples exhibiting dominant microporosity (DTP1), thus suggesting that micropore networks from Cape Range behave like soft or reference pores (*sensu* Xu and Payne, 2009). Indeed, such micropores, are located in between micrite calcite crystals, represent very flat pores and explain their low α_K signature. The 702 velocity-porosity relationships in carbonates with dominant microporosity have been commonly reported to be well modelled by the time average equation (Willy, 1951) and, as a 703 consequence, micropore networks have been assumed to match with reference pores 704 705 (Anselmetti and Eberli, 1999, Eberli et al., 2003; Weger et al., 2009). However, the high diversity of micrite microfabrics described from various microporous reservoirs has been 706 707 shown to result in a variety of acoustic behaviors, with pore networks ranging from soft (α_K from 0.05 to 0.15) to stiff (α_K from 0.15 to 0.25) domain (Fournier et al., 2011; Hairabian et 708 al., 2014, Fournier et al., 2018). In the present study, samples with dominant microporosity 709 710 (DPT1) exhibit a similar scattering in α_K with values ranging from 0.04 to 0.21. Such a scattering in α_K values may therefore result from a polyphase evolution of micrites or from 711 variable degrees of micrite transformation during diagenesis. Finally, the samples with 712 713 dominant intergranular macropores (DPT 3) show intermediate α_K values around 0.2-0.3, belonging to the stiff domain. This relative stiffness of the intergranular pore space in Cape 714 715 Range samples is plausibly related to the partial occlusion of such intergranular pores by development of sparry calcite cementation around grains that, on one hand, stiffens the solid, 716 717 mineral network and, on other hand, leads to the formation of relatively equant residual pores 718 in the center of the intergranular space, thus behaving acoustically as an intermediate case between micropores and moldic pores. 719

720 **5.2.2** Contribution of 3D CT Scan sections in underlining progressive diagenetic processes

A significant scattering of α_K values is found within the distinct DPT classes (Fig. 12) with standard deviation for DPT4 and DTP1 classes at 0.15 and 0.06 respectively. To better understand the origin of such a α_K variability, we used micro 3D CT-scan models (Fig. 15) in order to (i) detect the possible occurrence of pore type heterogeneities at the centimetre/plug scale, and (ii) to assess the possibility of subtler changes in the pore network. Thus, the CT 726 images allow revealed why samples A10 and A8, in spite of the occurrence of molds and vugs of mm to cm scale, present very low α_K values of 0.05 and 0.11 respectively (A10 and 727 A14 in Fig. 15). Open fractures and/or open stylolites therefore results in significantly 728 softened pore networks, shifting their α_K value to shift into the soft-pore domain 729 ($\alpha_K < 0.2$, black arrows in Fig. 12). This acoustic behavior reflect the very flat shape and low-730 stiffness of such discontinuities, known to strongly reduce Vp and Vs (e.g. Pyrak-Nolte et al., 731 1990; Matonti et al., 2015; Matonti et al. 2017). The samples exhibiting the highest or lowest 732 α_K values show very different pore network on CT-Scan. The samples with the lowest α_K 733 mostly display homogeneous micropore network (in grey, A8 in Fig. 15), with scarce 734 macropores. In contrast, the samples with the highest α_K values (around 0.5) contain only 735 736 large and/or rounded molds and vugs (in black, Tr5 and Tr6 in Fig. 15D), making them even stiffer compared to lower α_K samples in DPT4 class (i.e. $\alpha_K \approx 0.2$; Fig. 15). 737

For samples located at the transition from the soft to the stiff pore domains (i.e. DTP1 and 738 739 DTP4) ($\alpha_K \approx 0.2$), the CT-scan images exhibit similar patterns and proportions of micro and macropores, regardless of their DPT determined from thin section point counting (A12-13, 740 741 Tr3-6 in Fig. 15). These patterns result from similar dissolution processes affecting especially 742 the LBF in the Mandu (e.g. Fig. 6B) or grains and matrix in the Trealla (e.g. TR3 in Fig. 15), which formed a few small moldic/vuggy pores. These are better captured on the 3D CT-scan 743 data (black box, A13 in Fig. 15) compared to the quasi-2D thin sections. Moldic pore from 744 745 LBF leaching are flat, as a result of the original shape of these organisms, and may explain 746 their relatively low stiffness, thus low α_K signature (i.e. close to 0.2) for these DPT4 samples Additionally, the dissolution phase that led to the development of moldic pores and vugs 747 748 could have also, in an early stage, deeply restructured the micropore space into a microvacuolar network) as evidenced in the Urgonian limestones from Provence (Fournier et al., 749 2014, 2018). These grains are thus still competent but may already behave acoustically as 750

quasi-molds (blue arrow in Fig. 12). It is interesting to note that relatively low levels (less than 5-10%) of stiff macropores within a dominantly microporous pore network, as revealed by CT-scan analysis, are sufficient to induce a significant increase in α_K . Therefore, in the perspective of pore type discrimination from acoustic properties, such threshold effects should be considered. Finally, it appears that α_K variations are largely related to the dominant pore type in each sample and are able to capture pore network structures.

757

5.3 Combined effect of depositional and diagenetic processes on the evolution of acoustic properties of heterozoan and photozoan carbonates

Our research demonstrated significant differences in pore network architectures and petroacoustic responses (α_K) between the heterozoan carbonates from the Mandu/Tulki Limestones, likely related to temperate to sub-tropical environments, and the overlying photozoan sub-tropical to tropical Trealla Limestone, probably deposited in sub-tropical to tropical setting.

Two main causes can be invoked to explain these differences. Firstly, the drastic change in 766 mineralogy of the dominant primary carbonate producers (calcite in Mandu/Tulki formations, 767 aragonite in the Trealla formation) has driven two distinct patterns of diagenetic evolution. 768 The dominance of calcite producing organisms (LBF, red algae, etc) favoured the 769 development of microporosity, whereas the dominance of biota with less-stable aragonitic 770 skeleton favoured the development of molds and vugs, during minor phases of subaerial 771 exposure. As a consequence, moldic/vuggy photozoan limestones from the Trealla Limestone 772 773 exhibit a stiffer acoustic signature, for a given porosity compared to microporous Mandu/Tulki heterozoan limestones. 774

Secondly, the emersion and subsequent marine transgression terminating the Tulki Formationled to significant erosion of the Tulki limestones. This may well explain the apparent higher

777 burial depth indicated by the abrupt shift in compaction through pressure-solution (concavo-778 convex grain contacts and stylolites) between the Tulki and the overlying Trealla Limestone. This is why the shallow-buried Trealla have been able to preserve a significant part of 779 780 intergranular pores, which also prevented the development of stylolites and fractures after rock early embrittlement (La Bruna et al., 2020), therefore preserved from pore network 781 softening. Additionally, higher burial, associated with little intergranular early cementation, 782 783 characteristic of non-tropical ramp carbonates (e.g. James, 1997; Frank and James, 2017), favoured tight grain re-arrangement. This did not allow the preservation of large and 784 785 intergranular pores in the Mandu/Tulki, increasing the relative amount of microporosity over macroporosity. These results therefore suggest that typical heterozoan, non-tropical 786 787 carbonates, are likely to be less prone to develop high moldic macroporosity and stiff 788 acoustic signature compared to the photozoan carbonates when subject to meteoric 789 diagenesis.

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5.4 Importance of integrative approaches for 3D modelling of carbonate systems and reservoirs

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An important challenge of 3D modelling of carbonate systems and reservoirs is the 794 795 integration of geological concepts making it possible to produce models that are more realistic than those that would be obtained only from well and seismic data. The use of 796 797 forward modelling approaches revealed to be extremely relevant for building realistic models that can be shared by the various disciplines of carbonate exploration such as sedimentology, 798 stratigraphy, diagenesis, seismic, geomodeling and reservoir geology (Lanteaume et al., 799 800 2018). Moreover, such approaches allow sensitivity tests on stratigraphic and diagenetic scenarios, sequence stratigraphic concepts and controlling factors to be performed. The 801 approach used in the present work allowed establishing and quantifying the link between 802

carbonate producers, nature of dominant pore types and acoustic properties for a given
diagenetic history controlled by subaerial exposure and subsequent erosion. The
establishment of such links is fundamental in forward modeling since it makes it possible to
convert depositional facies into acoustic properties according to different diagenetic
scenarios.

Moreover, the 3D variability of the physical properties underlined at the plug scale from CTscan data have been shown to be significant, especially when microstructural discontinuities occurs. Capturing such a plug-scale variability has been shown to be of importance for the upscaling of acoustic properties at reservoir scale (Bailly et al., 2019b). In the future it would be interesting to perform full 3D analyses, trying to extract useful descriptors (such as continuity and tortuosity of the solid skeleton) from actual CT-scan 3D pore network in order to make direct linkage between acoustic and reservoir properties.

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816 VI. Conclusions

The present work provides novel results which demonstrate the importance of the combined role of sedimentary attributes (depositional texture and mineralogy) and the diagenetic processes when assessing variability and specificity of the petroacoustic signature of heterozoan and photozoan carbonates. Our multidisciplinary and multiscale study of Oligo-Miocene carbonate deposits from the Cape Range peninsula revealed various issues concerning the link between the nature of carbonate production, the subsequent diagenetic and pore type evolution and acoustic properties:

Heterozoan, non-tropical, carbonates from Mandu and Tulki formations, in the proximal
part of a mainly Miocene carbonate ramp, are characterized by dominant microporosity and

826 significant pressure-solution compaction features leading to reduced pore network stiffness
827 and – consequently – decreased wave velocities.

The sub-tropical to tropical carbonates from Trealla formation, displaying mainly moldic
and partially cemented intergranular pores, display stiffer pore network leading to higher
wave velocities.

These petroacoustic signatures are influenced by other diagenetic features, such as
fracture/stylolites and partial grain partial dissolution, which can be detected from combined
thin-section and 3D-CT-scan data.

The variation in diagenetic evolution found for each of the three formations investigated is primarily related to: (i) differences in the initial intergranular cementation and compaction level, (ii) change in the skeletal mineralogy of the dominant carbonate producers.

Finally, the resulting dominant pore type information has been accurately retrieved from indirect petroacoustic data by using the Equivalent Pore Aspect Ratio proxy (α_K). Additional CT-scan sections, revealed that α_K is controlled, to some extent, by the transitional evolution of pores between soft to stiff pore network, reflecting increasing dissolution. The relevance of the α_K proxy assessed in this geological context and others suggests promising applications to the abundant well log data in the offshore domain.

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- 851 **Figures captions:**
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Figure 1: A Location and latitude drift of the Cape Range area from the deposition of the 853 Mandu formation (green contour) and the Trealla formation (blue contour) and present day 854 855 location (black contour); B: Geological map showing the 4 main studied outcrops (blue-grey colour correspond to late Miocene Detrital/Eolian sandstone); (1: Shothole Canyon; 2: 856 857 Charles Knife Road; 3: Unnamed quarry 1; 4: Unnamed quarry 2). C: Conceptual diagram illustrating the dynamo-stratigraphic relationships between the three Oligo-Miocene 858 859 formations of the Cape Range region: The Mandu and Tulki Formations are partly coeval facies, separated from the overlying Trealla Formation by up-lift and erosion in the proximal 860 part of the system. 861

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Figure 2: Diagram illustrating the multidisciplinary geological and petrophysical protocol.

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Figure 3: A: Synthetic log section of the Cape Range Oligo-Miocene deposits (modified from
Collins et al., 2006). B-G: Outcrops photographs of the Mandu, Tulki and Trealla formations.

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Figure 4: Microphotographs of the 3 facies associations in Polarized light. A: Lower Mandu
(F1A) peloidal, foraminiferal wackestone with large and disc-shaped benthic-foraminifera
(Cycloclypeus); B: Quartz rich, peloidal and finely bioclastic packstone (F1B); C: Sharp
erosive contact between foraminiferal packstone and sparitic grainstone in the Tulki
Limestone (F2); D: Trealla packstone to grainstone dominated by recrystallized green algae
skeleton (F3); E: Trealla packstone facies dominated by encrusting red algae and corals (F3).

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Figure 5: A: Matrix micrite content variation through the Mandu-Tulki limestone. B: Calcitecontent of whole-rock samples along the Mandu-Tulki part of the sedimentary succession.

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Figure. 6: Microphotographs of the diagenetic features observed in the Mandu Fm (A-B), the Tulki Fm (C-F) and the Trealla Fm (G-H). A: compaction features: microstylolites (red arrow) and mode-1 fractures (black arrow); B: Partially dissolved LBF (lepidocyclinid); C: Concavo/convex grain contacts and iron oxides located especially in the intraskeletal space; D:Early isopachous fringe of calcite cement (ISC) around bioclasts and (SC) sparry calcite in intergranular position, filling the dissolved grains voids, along with possible evidence of pendulous cements; F: ISC cements showing iron oxide fringe at the boundary with SC; G: ZSC prismatic cement; H: Large moldic pore, originating from the dissolution of an aragonitic green algal skeleton.

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Figure. 7: Microphotographs of thin-sections under polarized-light and cathodoluminescence
showing the different cement phases. A-B: Mandu Fm, Dark blue (SC) calcite filling the
recrystallized micrite area along with the burial fractures (Mandu); C-D: Tulki Fm, ISC and
SC cements; E-F: Trealla Fm, showing ZSC and little developed SC cement.

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Figure 8: A: Outcrop photograph from the "unnamed quarry 2" outcrop. In yellow: the principal fault plane; in red: paleokarstic breccia at the boundary between the Tulki and the Trealla facies. B: Close-up on the paleokarst breccia presenting typical terra-rossa matrix. C: Close-up on the crosscutting relationship between vertical, fault related, karst deposits and the older horizontal karstic void filled with coarse calcite crystals, in the underlying Tulki Limestone; D-E: Microphotographs illustrating the pervasive dissolution phase affecting grains (D) and matrix (E) in the Tulki Limestone.

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- Figure 9: Carbon and oxygen stable-isotopes: A: δ^{13} C vs δ^{18} O crossplot (dashed green box is the range of Miocene isotopic values; Veizer et al., 1999); B: Log of the vertical variations of δ^{13} C through the entire sedimentary succession investigated.
- Figure 10: Vp porosity (A) and Vs porosity (B) crossplots with samples labelled as afunction of the depositional facies association.
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907 Figure 11: Dynamic Bulk modulus vs. porosity crossplots with samples labelled with the908 depositional facies association (A), and dominant pore type (B).

909 Figure 12: Crossplot of α_K values as a function of porosity. Samples are labelled as a function

910 of the dominant pore type (DPT) and facies association.

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Figure 13: Paragenetic sequences of the 3 formations highlighting the strong differences in 912 diagenesis history between Mandu-Tulki and Trealla formations leading to different final 913 petrophysical properties. Coloured arrows indicate the time of deposition for each formation 914 915 along time. A: Schematic burial depth curve for the 3 formations (depth variations for the Tulki Fm is tentative, and based on minimal burial depth needed for compaction features 916 observed); B-D: Individual paragenetic sequence presented in relative time scale. E-F: 917 Comparison of microstructures and diagenetic features of the Tulki Fm (E) and Trealla Fm 918 (F) formations showing the pronounced difference in terms of compaction and pore type. 919

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921 Figure 14: Conceptual block diagram models illustrating the difference in depositional922 systems between the Mandu-Tulki succession (A) and the Trealla succession (C).

923 Figure 15: Micro CT-scans cross-sections showing variable α_K values, highlighting the 924 relationship between α_K proxy and actual pore network structure.

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926 Table 1: A: Facies associations main characteristics B: Pore type classification

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Table 2: Table of laboratory petrophysical plugs measurements (and associated moduli and α_k computed values), CaCO₃ content and dominant pore type class from thin-sections scans. Average Vs is the average of the 2 perpendicular polarized S-wave velocities. Sample marked with a * are the one in which some macropore is larger than the average wavelength of Pwave used for the measurement.

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D



























Facies Association	TextureMain benthic foraminiferal taxaOther common allochems		Depositional evironment		
F1	Wackestone (F1A) - Packstone (F1B)	ckestone (F1A) - <i>Cycloclypeus</i> and Globigerinids, ackstone (F1B) <i>Lepidocyclina</i> Quartz		Lower oligophotic zone (F1A) - Mesophotic zone (F1B)	
F2	Packstone- Grainstone	Rotaliids, Encrusting forams, Milliolids	Echinoderms, Red Alage	Euphotic zone	
F3	Grainstone- Bafflestone Miliolids, Soritids, Alveolinids Red		Scleractinian, Green algae, Red algae	Euphotic zone (<10m)	

В

		THIN-SECTION SEMI-QUANTITATIVE ANALYSIS (POINT-COUNTING)							
		Dominant Micro/intercrystalline	Mixed Micro	Dominant macroporosity (>10µm)					
		porosity (<10μm)	Intergranular Mold Macroporosity pores dominate pores macropore space macrop						
IMAGING	No open stylolite / fracture	DPT1	DPT2	DPT3	DPT4				
CT-SCAN	Occurrence of open stylolite / fracture	DPT1S	DPT2S	DPT3S	х				

Plug	VP (m/s)	Average VS (m/s)	Helium	Bulk	Shear modulus	ακ	Calcite	Facies	Dominant
Name	(at 40MPa)	(40MPa)	Porosity (%)	modulus	(Gpa)		percentage (%)		pore type
				(GPa)					
14/10	6402	3338	0.62	69.38	29.60	0.413	100	F2	DTP4
/15-3									
14/10	5798	3114	5.61	52.66	24.67	0.218	100	F2	DTP3
/15-4									
14/10	5711	3076	5.90	50.73	24.00	0.196	100	F2	DTP2
/15-5	5040	2407	0.54	57.00	0.00	0.440	400	50	DTDD
14/10	5949	3187	2.51	57.09	26.54	0.110	100	F2	DIP2
/15-6	6454	2205	1.20	C1 72	20.02	0.000	100	52	DTD2
	6154	3305	1.36	61.72	28.93	0.089	100	FZ	DIP3
/15-/	5776	2142	1.40	F1 07	25.41	0.020	100	52	
14/10	5770	5145	1.49	51.97	25.41	0.039	100	FZ	DTP4
/15-0	5082	2220	2.02	56.96	27.10	0 1 2 7	100	E2	
/15-9	3983	5250	5.05	50.80	27.10	0.157	100	ΓZ	DTF4
14/10	6159	3275	1.22	62.02	28.14	0.082	100	F2	DTP3
/15-									
10									
14/10	6192	3289	0.99	63.55	28.72	0.080	100	F2	DTP3
/15-									
11									
Msh	4422	2503	9.46	27.26	15.25	0.093	97.05	F1A	DTP1
A7									
Msh	5068	2882	6.25	37.01	21.03	0.096	94.4	F1A	DTP1
A8									
Msh	4074	2425	12.54	20.66	13.87	0.092	88.05	F1A	DTP1
A9									
Msh	3028	2309	21.25	4.39	11.34	0.056	91	F1B	DTP1S
A10 *									
Msh	5335	2979	8.92	41.01	21.88	0.179	98.05	F1B	DTP1
A11									

Msh A12	4510	2636	18.82	24.31	15.26	0.197	ND	F1B	DTP1
Msh A13	4500	2620	16.55	25.08	15.51	0.170	99.9	F1B	DTP1
Msh A13bi s	4596	2610	18.24	26.72	15.12	0.216	99.9	F1B	DTP1
Msh A14	4815	2804	10.01	30.98	19.16	0.117	100	F2	DTP3S
Msh A15bi s	2919	1762	28.32	8.40	5.95	0.120	86.95	F1A	DTP1
Msh A16	2352	1367	35.87	5.21	3.20	0.126	85.85	F1A	DTP1
Msh A17	3077	1818	9.90	10.82	7.07	0.044	ND	F2	DTP1
Msh A18	5807	3146	4.70	52.87	25.51	0.171	100	F2	DTP2
Msh A19	5843	3142	6.25	52.98	24.92	0.268	100	F2	DTP4
Tr1	6086	3212	4.71	59.98	26.57	0.443	100	F3	DTP4
Tr3	4568	2595	19.79	25.94	14.69	0.234	100	F3	DTP4
Tr4 *	6017	3223	7.54	55.85	25.95	0.578	100	F3	DTP4
Tr5 *	5739	3010	8.41	51.50	22.38	0.388	100	F3	DTP4
Tr5bis	5731	3110	8.09	49.47	23.98	0.291	100	F3	DTP4
Tr6	6142	3247	4.85	60.53	26.97	0.535	100	F3	DTP4
Tr7 *	4727	2711	16.91	28.35	16.60	0.211	100	F3	DTP4
Tr7bis	5266	2921	12.05	39.03	20.35	0.243	100	F3	DTP4
TrW1	6006	3218	5.47	56.55	26.29	0.329	100	F3	DTP3
TrW1 bis	5837	3157	6.91	51.47	24.70	0.270	100	F3	DTP4
Tuch1	6064	3198	2.41	60.55	26.77	0.159	100	F2	DTP2

Tuch1	6265	3299	1.15	65.42	28.77	0.143	100	F2	DTP2
bis									
Tuch2	6168	3193	1.52	64.67	26.95	0.175	100	F2	DTP1
Tuch3	6060	3196	1.22	61.17	27.05	0.073	100	F2	DTP1
Tuf2	5815	3134	6.91	52.20	24.76	0.292	100	F2	DTP4
Tuf2bi	5997	3220	4.87	56.78	26.59	0.277	100	F2	DTP4
S									