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Seagrass-related carbonate ramp development at the front of a fan delta (Burdigalian, New Caledonia): Insights into mixed carbonate-siliciclastic environments

Tournadour Elsa ^{1,2,*}, Fournier F. ³, Etienne S. ^{1,2}, Collot Julien ², Maurizot P. ², Patriat Martin ⁴, Sevin B. ², Morgans H.E.G. ⁵, Martin-Garin B. ³, Braga J.C. ⁶

¹ Adecap Technopole, 1 bis Rue Berthelot, Doniambo, B.P. 2384-98 846, Nouméa, New Caledonia

² Service Géologique de Nouvelle-Calédonie, DIMENC, B.P. 465, 98845, Nouméa, New Caledonia

³ Aix Marseille Univ, CNRS, IRD, INRAE, Coll France, CEREGE, Case 67, 3, 13331, Marseille Cedex 03, France

⁴ IFREMER, Unité Géosciences Marines, 29280, Plouzané, France

⁵ GNS Sciences, PO Box 30368, Lower Hutt, 5040, New Zealand

⁶ Universidad de Granada, Departamento de Estratigrafía y Paleontología, Avenida de La Fuente Nueva, 18071, Granada, Spain

* Corresponding author : Elsa Tournadour, email address : elsa.tournadour@gouv.nc

Abstract :

Past shallow-water carbonate environments of the main island of New Caledonia (NC) have been subject to high terrigenous influx derived from the erosion of ultramafic obducted nappes and are therefore a relevant case study for characterizing neritic carbonate production in mixed carbonate-siliciclastic systems under a tropical climate. More particularly, we focused on Burdigalian carbonate sedimentary records cropping out on the western coast of NC, in the Népoui area. Based on a comprehensive sedimentological study of cores and outcrops, we established a new depositional model of an alluvial fan to carbonate ramp transition. Shallow-water (euphotic) carbonate production was dominated by seagrass-related biota and corals derived from small-sized bioconstructions. Extensional tectonics and associated normal faulting, driven by post-obduction isostatic rebound, favored carbonate ramp aggradation and preservation. The carbonate ramp was incised by conglomerate-filled terminal distributary channels, which indicate that terrigenous inputs remained significant during marine transgression and did not inhibit the development of seagrass and scleractinian carbonate factories. Phytal substrates induced by seagrass and/or macro-algae seafloor colonization strongly controlled the nature of carbonate production and promoted the accumulation of foraminiferal-coralline algal sediments. Seagrass development at the front of the fan delta is interpreted to have controlled the preservation of a diverse and significant carbonate production by reducing water turbidity and by limiting the risk of suffocation for filter-feeding biota in such a high terrigenous influx setting.

Highlights

► Extensive development of seagrass carbonates occurs at the front of tropical fan delta. ► Seagrass meadows development favored carbonate production and preservation. ► The long-term preservation of the depositional system is due to the post-obduction subsidence regime.

Keywords : Miocene, seagrass, tropical carbonates, terrigenous inputs, alluvial fan, facies association, New Caledonia

43 **1. INTRODUCTION**

44 Mixed carbonate-siliciclastic depositional systems are characterized by a high vertical and
45 lateral variability in facies and lithologies, which results from the interplay between carbonate
46 production and terrigenous supply. Such complexity makes difficult the sequence stratigraphic
47 interpretation of mixed systems (Zeller et al., 2015) and complicates the prediction of reservoir
48 architectures from subsurface data (McNeill et al., 2004).

49 Mixed systems have been commonly subdivided into temporal phases of either dominant
50 carbonate or siliciclastic deposition by applying the concepts of reciprocal sedimentation
51 (Wilson, 1967). According to this model, siliciclastic deposits prevail during low sea level
52 periods whereas carbonate facies dominate during transgressions and highstands. However, this
53 model, often used in ancient case studies (e.g. Permian Capitan margin: Silver & Todd, 1969;
54 Kerans & Tinker, 1999) has been shown to be inadequate for describing several quaternary (e.g.
55 Great Barrier Reef: Page et al., 2003; Harper et al., 2015; New Caledonia: Le Roy et al., 2019) and
56 ancient mixed carbonate-siliciclastic systems (e.g. Late Jurassic northeastern France: Olivier et
57 al., 2004; Jurassic-Cretaceous Neuquén basin, Argentina: Zeller et al., 2015). The deviations from

58 the reciprocal sedimentation model may be related to increased siliciclastic inputs during
59 transgression as a result of regional climate changes, reworking and dispersing of previously
60 deposited terrigenous sediments (Harper et al., 2015) or steeper fluvial profile compared to
61 shoreline trajectory (Catuneanu, 2019). In addition, continued high carbonate production during
62 sea level lowstands and/or periods of high terrigenous inputs go against the applicability of
63 reciprocal sedimentation concepts (Zeller et al., 2015). Carbonate production and the nature of
64 carbonate producers are therefore a key parameter to be considered for the prediction of
65 depositional models and stratigraphic architectures in mixed systems. Moreover such
66 predictions are complicated by the interplay between carbonate production and siliciclastic
67 inputs. Terrigenous influx is known to be a significant factor controlling carbonate production
68 since it leads to increased water turbidity, thus reducing available light for light-dependent
69 organisms and to suffocation of filter-feeding biota due to high sedimentation regimes (e.g.
70 Rogers 1990; Fabricius 2005). Many studies have shown that prolonged exposure to terrigenous
71 discharge from rivers has a negative impact on shallow-water carbonate production, particularly
72 in coral reefs (Rogers 1990; Van Katwijk et al. 1993; Fabricius 2005). In contrast, such a negative
73 impact of terrigenous inputs on neritic carbonate production has been contradicted in different
74 Quaternary and ancient tropical carbonate systems from the Indo-Pacific (e.g. Wilson, 2005;
75 Perry et al., 2012; Novak and Renema, 2015; Santodomingo et al., 2015) and Mediterranean
76 domains (e.g. Corsica: Tomassetti et al., 2013; Sardinia: Vigorito et al., 2012; Ligurian Alps:
77 Brandano et al., 2015; Turkey: Bassant et al., 2005), and many examples of barrier-reef systems
78 developing on terrigenous shelves were documented in modern environments (The Great
79 Barrier Reef in Australia: e.g. Belperio, 1983; Orpin et al., 2004; Puerto-Rico: Ryan-Mishkin et al.,
80 2009; Belize shelf: Mc Neill et al., 2010; Gulf of Mexico: Hernandez-Arana et al., 2015). These
81 numerous examples have demonstrated the great capacity of photozoan, coralgall production to
82 adapt to strong terrigenous inputs. However, even though seagrass-related biota has been
83 reported in various mixed carbonate-siliciclastic systems (e.g. Reuter et al., 2011; Sola et al.,
84 2013; Beavington-Penney et al., 2004), relatively low attention has been paid on the resilience of

85 seagrass carbonate factories to strong terrigenous discharges (Mateu-Vicens et al., 2012).
86 Additionnaly, although Cenozoic seagrass deposits are well known from the Caribbean (Eva,
87 1980), the Mediterranean (e.g. Mateu-Vicens et al., 2008; Brandano et al., 2009; Bassi and
88 Nebelsick, 2010; Sola et al., 2013; Tomassetti et al., 2016) to the Indo-Pacific (e.g. Chaproniere,
89 1975; Beavington-Penney et al., 2004; Reuter et al., 2011; Haig et al., 2020) areas, complete
90 depositional models of seagrass-influenced carbonate systems subject to high terrigenous
91 supplies are lacking.

92
93 The Upper Népoui Formation (Burdigalian), cropping out on the western margin of New
94 Caledonia (South-West Pacific), is a mixed shallow-marine depositional system where
95 continuous terrigenous inputs derived from post-obduction erosion of ultramafic nappes were
96 supplied into shallow-water ramp environments where a significant seagrass-related, and
97 scleractinian carbonate production occurred (Maurizot et al., 2016). Such spatially well-
98 constrained outcrops thus represent a relevant case study for understanding the ecological and
99 physical processes that control the architecture and evolution of tropical, seagrass-dominated
100 carbonate systems which are subject to high terrigenous supplies.

101 A comprehensive sedimentological study of the Burdigalian mixed system has been performed,
102 based on newly acquired core data as well as new outcrop descriptions. On the basis of the
103 detailed geological mapping and logging of sedimentary units, coupled with micro- and macro-
104 scale petrographic and sedimentological characterisation of the carbonate and terrigenous
105 deposits, the present work aims at: 1) constructing an original depositional model of the
106 transition from an alluvial-fan to a seagrass-dominated carbonate ramp, 2) assessing the role of
107 potential controlling factors such as relative sea level fluctuations, nutrient supplies and water
108 turbidity on the depositional architecture of mixed carbonate-siliciclastic systems, and 3)
109 discussing the significance of seagrass-related carbonate factories in tropical environments
110 subject to high-terrigenous influx.

111 The resulting depositional architecture constitutes a valid and original model for potential
112 hydrocarbon reservoirs in mixed carbonate-siliciclastic ramps forming in tropical setting.

113

114 **2. GEOLOGICAL SETTING**

115

116 **2.1 Geodynamic framework of the New Caledonian western margin**

117 The New Caledonian main island, Grande Terre, is located in the South West Pacific (Fig.1), at
118 the north-eastern tip of Zealandia, a large and mostly submerged piece of thinned continental
119 crust inherited from the cretaceous break-up of eastern Gondwana (Mortimer et al., 2017).
120 Indeed, from Late Cretaceous to Paleocene, a rifting phase followed by seafloor spreading in
121 Tasman Sea led to the isolation of Zealandia (Gaina et al., 1998). During the Eocene, New
122 Caledonia underwent a convergence phase that led to the NE-SW emplacement of various
123 nappes, including the prominent mantellic Peridotite Nappe (Paris, 1981; Cluzel et al., 1994;
124 Aitchison et al., 1995; Cluzel et al., 2001; Cluzel et al., 2012). During obduction, the New
125 Caledonia Basin, which is located west of Grande Terre, reacted as an asymmetrical flexural
126 basin, and is now filled by 6 km-thick post-obduction sediments (Collot et al., 2008).

127

128 The tectonic and sedimentary evolution of Grande Terre during the post-obduction period
129 (Oligocene-present) remains debated. Most authors agree that this period is characterized by an
130 uplift of Grande Terre driven by post-obduction isostatic rebound (Moretti & Turcotte 1985,
131 Lagabrielle & Chauvet, 2008, Collot et al., 2017). This uplift is interpreted as being at the origin
132 of the collapse of both the western and eastern margins through large scale normal faults
133 together with the development of a deepwater fold and thrust belt at the foot of the western
134 margin of New Caledonia (Lagabrielle & Chauvet, 2008, Collot et al., 2017). However, based on
135 the subduction-related geochemical signatures of the Koum and St Louis granitoids intruding the
136 Peridotite Nappe, Cluzel et al. (2005) and Paquette and Cluzel (2007) suggested that a short-

137 lived east-dipping Oligocene subduction of the New Caledonia Basin occurred beneath Grande
138 Terre.

139
140 More recently, since the Plio-Quaternary, Grande Terre has been also affected by antithetic
141 vertical motions between its western and southeastern margins. Indeed, based on drillings
142 performed both on fringing reefs and on the barrier reef, Cabioch et al. (1999) suggested that a
143 subsidence of the western margin of New Caledonia allowed the aggradation of the barrier reef
144 and the associated lagoon. Montaggioni et al. (2011) estimated this subsidence to be higher than
145 $0.08 \text{ mm} \cdot \text{yr}^{-1}$ since 1 Ma ago. Conversely, along the southeastern margin of New Caledonia, a ca.
146 10 m uplift of reefs is observed in the area of Yaté and Pines Island (Cabioch et al., 1996). These
147 reef uplifting is thought to be the result of the flexure of the lithospheric plate entering into the
148 New Hebrides subduction (Dubois et al., 1974; Lafoy et al., 1995).

149

150 **2.2 Geological setting of the Népoui area**

151 The mixed siliciclastic carbonate deposits of the Népoui Group crop out along the western
152 margin of New Caledonia, in front of the peridotite massifs of Kopéto and Boulinda. They
153 unconformably rest on both the Népoui Eocene Flysch and the Poya Terrane (Figs.2A and B).
154 The Népoui Flysch, dated from the late Eocene by micropaleontologic analyses, is considered as
155 a syn-obduction formation based on its mafic lithological composition and its deformed state
156 (Cluzel et al., 1998). It constitutes the youngest formation located beneath the allochthonous
157 Peridotite Nappe, which would constrains the oldest possible age of the end of obduction at 34
158 Ma. The Poya Terrane comprises mid ocean ridge basalts and abyssal sediments, dated from the
159 Late Cretaceous to the latest Paleocene based on radiolarian fauna, and is situated at the base of
160 the Peridotite Nappe (Paris, 1981; Cluzel et al., 1997) (Figs.2A and 2B).

161

162 First descriptions of the Népoui Group considered a relatively simple vertical succession
163 involving a lower fluvio-deltaic unit, named the Pindaï Conglomerates, overlain by shallow

164 marine limestones, initially attributed to the Early to Middle Miocene (Gubler & Pomeyrol,
165 1948; Coudray, 1975; Paris, 1981) (Figs.2A and B). More recently, Maurizot et al. (2016)
166 discovered a new limestone unit below the Pindaï Conglomerates, on the Chapeau Chinois
167 outcrops and in the 103 m deep S4 well (see location on Fig. 2), which reached the unconformity
168 separating post-obduction miocene deposits from the syn-obduction eocene Népoui Flysch
169 (Fig.2B and Fig.3). These authors consequently subdivided the mixed system of Népoui into
170 three formations and restricted the lower limestone formation to the Aquitanian and the upper
171 limestone formation to the Burdigalian based on six strontium isotope measurements performed
172 on preserved Mg-calcite from echinoid spines, combined with benthic and planktonic
173 foraminiferal biostratigraphy (Fig. 3). The lower formation (Lower Népoui Formation) is a
174 dominantly carbonate interval composed of three members, from base to top: the A1 Chapeau
175 Chinois Member, the A2 Operculina Green Sands Member and the A3 Xuudhen Limestone
176 Member (Fig. 3). Above, the strictly siliciclastic, fluvio-deltaic Pindaï Conglomerate Formation, is
177 assigned to the Burdigalian, yet no direct biostratigraphic or radiometric datings were
178 performed on that interval. The upper formation (Upper Népoui Formation) is made of mixed
179 carbonate-siliciclastic, shallow water deposits, Burdigalian in age, which are subdivided into two
180 members: the Wharf Member and the Népü Limestone Member (Fig. 3). In this paper, we argue
181 that the Pindaï Conglomerates are part of the Upper Népoui Formation, similarly to the
182 lithostratigraphic framework proposed by Sevin et al. (2014). Following this scheme, the Upper
183 Népoui Formation comprises the B1 Pindaï Conglomerate Member, the B2 Wharf Member and
184 the B3 Népü Limestone Member (Fig. 3). A dominant seagrass-related carbonate production
185 prevailed in the euphotic area of the Aquitanian and Burdigalian ramps (Maurizot et al., 2016),
186 even though scattered scleractinian colonies are common in both systems.

187
188 Finally, the post-Burdigalian Muéo Formation unconformably rests on the Népoui Upper
189 Formation on the Pindaï and Muéo peninsulas (Fig.2A). It primarily comprises extensive lateritic
190 red muds with large silicified blocks and reworked lateritic ferricretes (Paris, 1981). In former

191 studies, this formation was correlated with the Pindaï conglomerates, which led to some degree
192 of confusion (Paris, 1981).

193

194 In the offshore domain, seismic profiles revealed a 200 km along-margin and 50 km into the
195 basin deepwater fold and thrust belt in front of the Népoui area, interpreted as being gravity
196 driven (Rigolot & Pelletier, 1988; Collot et al., 2017) (Fig.1). Faults associated with this failure
197 are deeply rooted in the New Caledonia Basin (ca. 5 km depth) and are interpreted as being in
198 continuity with major detachments onshore that is in agreement with the extensional tectonic
199 model suggested by Lagabrielle & Chauvet (2008).

200 3. DATA AND METHODS

201 Outcrops of the Népoui Group are restricted to an area of about 100 km² that includes the
202 Pindaï, Muéo, Nékoró, and Béco peninsulas, as well as the islands and islets of Grimault, Didot
203 and Longue (Fig.4). This study focused on the Burdigalian Upper Nepoui Formation (Fig.3). The
204 overall stratigraphic succession is 80 m-thick and is slightly tilted (< 10°) towards the SSW, i.e.
205 towards the present day lagoon and Quaternary barrier reef (Fig.2). The most recent study
206 dedicated to the Népoui Group was based on the description of a 103 m deep cored well, S4, in
207 combination with its stratigraphic continuity exposed on the outcrops of the Chapeau Chinois
208 and Wharf areas (Maurizot et al., 2016) (Fig. 4). In order to improve the architectural
209 characterization of the Upper Népoui Formation (Burdigalian), the existing dataset was
210 complemented by a new, 170 m-long cored well, S2 (Fig. 4). The latter is located only a few
211 kilometers away from the main outcrops of the Pindaï peninsula and reveals the spatial
212 variability system in a proximal to distal direction, following a NNE-SSW transect. Additionally,
213 we included nine shallower drilled boreholes (20 to 30 m long) and detailed sedimentological
214 descriptions of coastline outcrops of the Pindaï and Beco peninsula and of the Grimault, Didot
215 and Longue islets, in order to constrain the depositional models and the stratigraphic
216 architectures (Fig. 4). Based on this new dataset, we performed a detailed sedimentological
217 analysis at the 1:50 scale using a classical “facies association” approach (eg. Allen, 1983; Miall,
218 1985; Posamentier & Walker, 2006). The outcrops and S2 core data were investigated by logging
219 7 key sections, which were summarized in 1:500 scale synthetic logs presented in Fig. 18.
220 Depositional facies were defined based on sedimentary structures, depositional textures and
221 biological composition identified from outcrops, cores and thin-sections. A total of 100 thin-
222 sections from the Upper Népoui formation were used for microfacies characterization (texture,
223 skeletal composition, diagenetic features). Additionally, 22 thin-sections were point-counted by
224 using the JMicroVision software on the basis of 300 points to quantify the bioclastic composition
225 of the sand-grained skeletal fraction of carbonates. Facies were grouped into facies associations,
226 which were interpreted in terms of depositional environments. Finally, the lateral and vertical

227 stratigraphic distribution of facies associations allowed a well-constrained depositional model to
228 be proposed.

Journal Pre-proof

229 4. RESULTS

230 4.1 Facies association analysis

231 The Upper Népoui Formation is a lithologically heterogeneous and sedimentologically complex
232 succession. The spatial and vertical distribution of the sedimentary facies, together with their
233 stacking patterns further permits to gather them into eight facies associations that are thought
234 to reflect distinctive depositional environments (Fig. 5). We identify continental deposits (FA1,
235 FA2 and FA3) within the B1 Member of the Pindaï Conglomerates, shallow-marine sandstone
236 deposits (FA4 and FA5) for the B2 Wharf Member and shallow-water carbonate deposits (FA6,
237 FA7, FA8) typifying the B3 Népü Limestone Member. These facies associations are detailed in
238 the following sections.

239

240 ***FA1- Proximal alluvial fan (Fig. 6)***

241 Description: in the northeastern part of the Pindaï Peninsula and on the Muéo Peninsula along
242 the “*piste de roulage*” outcrops (see location Fig. 4), the Pindaï Conglomerates display well-
243 rounded, moderately- to poorly-sorted pebbles in a coarse- to fine-grained sandy matrix (Fig.
244 6A). These polygenic conglomerates are composed of mafic and ultramafic clasts (serpentinites,
245 serpentinitised peridotites, dolerites, gabbros) as well as ferricrete clasts. They can reach 1 m in
246 diameter but their average size is 30-40 cm (Fig. 6A, B and C). In the western part of the Muéo
247 Peninsula, these conglomerates rest unconformably on the Népoui Eocene Flysch (Fig. 6E and
248 6F). Two fining upward conglomeratic to sandy successions are here exposed. They comprise
249 matrix-supported, disorganized to normally graded conglomerates with erosive base, that
250 evolve vertically towards gravelly sandstones with faint trough cross-bedding highlighted by
251 gravels and pebbles (Fig. 6D and 6F). Although pebble orientation is in most places chaotic, a
252 preferential north-dipping clast imbrication is observed, suggesting a north to south dominant
253 transport. This orientation is confirmed by the presence of N20° oriented groove marks at the
254 base of the second fining upward succession.

255
256 Interpretation: within the conglomeratic facies, the matrix-supported fabric, poor sorting of
257 clasts combined with the apparent lack of tractive sedimentary structures are consistent with
258 subaerial debris flow processes (Blair & McPherson 1994; Blair 1999). In turn, more structured
259 conglomerates and overlying cross-bedded gravelly sandstones are consistent with bedload
260 transport and tractive deposition by high density flows, and would possibly correspond to
261 braided-river barforms (Bluck, 1979; Miall, 1985). We therefore interpret this facies association
262 as recording deposition within the proximal part of an alluvial fan, where surficial braided
263 distributary channels can develop on top of alluvial fan debris flow lobes (e.g. sheetflood
264 dominated alluvial-fan model of Blair & McPherson, 1994). The lack of preserved floodplain
265 deposits is an additional evidence to favor an alluvial fan setting over a fluvial channel belt. In
266 addition, the lack of angular clasts in these facies is of note. Here, the subrounded to rounded
267 character of boulders and pebbles does not appear to be directly linked with flow transport
268 distance. Indeed, weathering of peridotites under a humid tropical climate can promote the *in*
269 *situ* formation of rounded blocks in saprolite layers (Trescases, 1975) (Fig. 6B). However, in the
270 absence of valuable geometrical constraints due to outcrop discontinuity, an alternative
271 interpretation where structureless conglomerates represent the basal part of channel-fills rather
272 than unconfined debris-flow lobes cannot be entirely ruled out.

273

274 ***FA2- Distal alluvial fan*** (Fig. 7)

275 Description: in the East of the Pindaï Peninsula, the basal part of the Wharf outcrops (see
276 location Fig. 4) exhibits a fining and thinning upward succession whose base consists in pebble-
277 cobble (2 to 15 cm in diameter) oblique bedded conglomerates, in places with pebbles preserved
278 as residual vugs (“ghost pebbles”), within a coarse-grained sandstone matrix with floating
279 granules (Fig. 7A and 7B). Several decimeter-thick intervals of normally graded, coarse-grained
280 trough cross-bedded sandstones with granules overlie these conglomerates. The whole
281 succession is organized as meter-scale oblique bedsets with dips of around 15–20° toward the

282 SSW (Fig. 7B). The succession is topped by red to greenish-red medium to fine-grained trough
283 cross-bedded sandstones (Fig. 7C) capped by a traceable surface highlighted by numerous roots
284 and iron oxide concretions (Fig. 7A and 7D).

285
286 *Interpretation:* similarly to FA1, oblique bedded conglomerates can be interpreted as sand-
287 gravel bars formed by the sediment discharge of a unidirectional high capacity flow (Allen, 1983;
288 Miall, 1985). Trough cross-bedded sandstones are consistent with fluvial settings and are
289 typically assigned to three dimensional dunes (Allen, 1963; Allen, 1983) which may be
290 interpreted to develop as frontal splays of alluvial fans. In our case, the SSW orientation of
291 bedsets, parallel to the known general transport direction, together with the grading of
292 sandstones points to a downstream accretion of deposits at the mouth of a stream channel
293 alluvial fan rather than lateral accretion within a river channel belt. At the top of the succession,
294 the decrease in grain size, increase in plant debris, root traces and iron concretions can be
295 interpreted as a planar abandonment surface and are consistent with paleosol development.
296 This facies association is therefore interpreted as a recording deposition within the distal part of
297 alluvial fans.

298

299 ***FA3- Fluvial channel-fills (Fig. 8)***

300 *Description:* at the north easternmost extremity of the Didot Islet (see location Fig.4), an isolated,
301 erosive based, 5 m-thick conglomerate succession is exposed (Fig. 8A and 8B: channel 1). It
302 comprises polygenic, mafic and ultramafic, 5 to 6 cm in diameter pebbles within a coarse- to
303 very coarse-grained sandy matrix organized in graded beds with erosional trough-shaped bases,
304 and meter-scale oblique sets, with a dominant SSE orientation (Fig. 8C). These deposits contain
305 numerous and large silicified wood fragments (Fig. 8D) including trunks, up to 1 m long.
306 Laterally, only a few tens of meters westwards, a similar 10 m-thick succession occurs (channel
307 2: Fig.8A). Both successions are overlain by an alternation of marly deposits and centimeter-

308 thick, fine- to very fine-grained calcareous sandstones with undifferentiated bioclasts and
309 foraminifers (Fig. 8A).

310

311 *Interpretation:* these well-structured conglomerates are interpreted as resulting from bedload
312 transport, most likely within a coarse braided river system (Allen, 1963; Miall, 1977). Indeed,
313 within such isolated channel-fills, individual trough cross-bedded sandstones record distinct
314 phases of the channel fill. These small scale channel fills are interpreted to be deposits in fluvial
315 to near-shore environments possibly as the downstream evolution of alluvial fan environments
316 (FA1 and FA2).

317

318 ***FA4- Tidal sands (Fig. 9)***

319 *Description:* in the south of Longue Island (see location Fig.4), a 30 m-thick succession of very
320 coarse to medium-grained cross-bedded sandstones, typically organized into sigmoidal
321 megaripples, is exposed (Fig. 9A and 9D). These decimeter-thick sigmoidal bedsets extend over
322 2 to 5 m and show a 15° to 30° inclination of individual sets. Measurements of oblique set
323 apparent dips reveal a main SE oriented progradation with subordinate orientations toward the
324 W, NE, and E (Fig.18). Sigmoidal lamina sets can be locally erosional and commonly show cyclic
325 increases and decreases in thickness within the same bed. Mud drapes, mud clasts, opposite
326 ripples and bioturbation are present even though relatively scarce, possibly due an overall low
327 fraction of mud-sized sediments. However, opposite megaripples (i.e. herring-bones) occur (Fig.
328 9B). In places, small-scale trough-cross ripple laminations are observed between bedsets.
329 Frequent gravel lags and centimeter to decameter scale isolated lenses of well-rounded
330 ultramafic pebbles within a calcareous muddy matrix are observed (Fig. 9C). These sandstones
331 are primarily made of altered serpentinite grains and ferricrete fragments. Grains are coated
332 with a fringing marine cement and separated from each other by drusy calcite cements (Fig. 9E).

333

334 *Interpretation:* The common preservation of alternating thicker and thinner sigmoidal foresets,
335 erosional (reactivation) bounding surfaces, together with herring-bone cross-bedding, strongly
336 suggests a tidal influence (spring-neap tide bundling). In addition, the apparent lack of storm
337 and wave-related sedimentary features (ie. HCS, SCS, oscillating ripple laminations) point to the
338 fact that the depositional environment is not significantly influenced by oscillating currents. The
339 weak preservation of mud drapes and the lack of bioturbation could be seen as evidencing rapid
340 bedform deposition by sustained, high-energy currents that reworked evidence for depositional
341 pauses (eg. Willis et al., 1999). Consequently, cross-bedded sandstones are therefore thought to
342 correspond to tidally influenced shallow marine sandstones, possibly conforming to tidal sand
343 bars in an estuarine environment (Belderson et al., 1982; Reynaud & Dalrymple, 2012; Snedden
344 & Dalrymple, 1999; Snedden, et al., 1999), or, alternatively, to distributary mouth bars in a tide-
345 influenced delta front.

346

347 ***FA5- Subtidal sands (Fig. 10)***

348 *Description :* in the west of the Pindai Peninsula, at the nautical base outcrops, as well as on the
349 Didot Islet and the Beco Peninsula (see location Fig.4), fine to very fine-grained, tabular cross-
350 bedded, typically bioturbated sandstones are exposed (Fig. 10A and 10D). The direction of
351 apparent progradation is dominantly towards the SSW, yet a few measurements were found to
352 be towards the E and NE (Fig. 14). Where grain size is finer, cross bedding is replaced by parallel
353 laminations and isolated unidirectional ripples (Fig. 10B). Bioturbation is also more pronounced
354 and trace fossils are more identifiable. Burrows include *Thalassinoides* and *Dactyloidites* (Fig.
355 10C). Additionally, numerous mafic and ultramafic granules are present, typically as lags or
356 within burrow infills. In S2 well, cross-bedded and bioturbated sandstones exhibit a scarce and
357 poorly diversified foraminiferal association dominated by poorly preserved *Miogypsina* and
358 *Sorites*.

359

360 *Interpretation:* these sandstone facies are interpreted as sandy subtidal deposits downslope of
361 FA-4 tidal sands. Their finer grain size, the ichnofacies (*Thalassinoides-Dactyloidites*) and the
362 lack of evidence for clear tidal features is consistent with deposition below the low tide limit. In
363 addition, while *Thalassinoides* burrows may characterize diverse depositional settings from
364 shallow to deep water environments (Bromley, 1996), *Dactyloidites* is usually considered to be
365 restricted to sandy substrate in very shallow, shoreface environments (Uchman & Perversler,
366 2007). Finally, the scarcity and poor preservation of benthic foraminifers suggest these are not
367 *in situ* but likely derived from neighboring euphotic to upper mesophotic environments (e.g.
368 Hallock & Glenn, 1986; Pomar et al., 2014).

369

370 ***FA6- Subtidal, seagrass-related, low-angle carbonate ramp (Fig. 11)***

371 *Description:* outcrops from the B3 Népü Limestone Member are dominated by massive and
372 chalky carbonates, which are commonly interbedded with centimeter to meter-thick intervals of
373 ultramafic cobbles and boulders (Fig. 11D). Three dominant carbonate depositional facies have
374 been identified within the FA-6 facies association: 1) FA6-1: Foraminiferal-coralline algal
375 packstone-grainstones (Facies FB2 in Maurizot et al., 2016), 2) FA6-2: Scleractinian floatstone
376 with a foraminiferal-coralline algal packstone-grainstone matrix (Facies FB1 in Maurizot et al.,
377 2016), and 3) FA6-3: Scleractinian boundstones.

378 Foraminiferal-coralline algal packstone-grainstones (FA6-1) are organized into decimeter to
379 meter-thick beds (Fig. 11B). They are dominated by benthic foraminifera and coralline algae,
380 and include small proportions of mollusks and echinoderm fragments. Scleractinians may be
381 present as sparse gravel to cobble-sized coral fragments but do not contribute significantly to
382 the sand-sized skeletal fraction. It is likely that part of the sandy fraction deriving from corals is
383 present as dissolved grains and has therefore been counted as indeterminate elements (<7 %).
384 Vertical calcitic tubes of shipworm (teredinid) bivalves, 1 to 2 cm in diameter and 10 to 30 cm in
385 length, may be common in a few beds (Fig. 11C). Packstones and grainstones exhibit very similar
386 biological composition (Tables 1 and 2, Fig. 12) with high proportions of benthic foraminifers

387 (52 % and 46 % respectively), articulated coralline algae (13 % and 16 %) and encrusting
388 coralline algae (17 % and 25 %). In both textures, the foraminiferal assemblage is dominated by
389 benthic foraminifers which are known as common epiphytes (e.g. Chaproniere, 1975; Hallock et
390 al., 2006; Pomar et al., 2014; Reich et al., 2014) such as the miliolid *Austrotrillina*, soritids (Fig.
391 13A, B, C, F), peneroplids, and calcarinids (*Neorotalia*) (Fig. 13D), which represent, in average,
392 30 % (packstones) and 27 % (grainstones) of the sand-sized bioclastic grains. Victoriellid
393 encrusting foraminifers (Fig. 13E) are common (7 % in average but may reach up to 47 % of the
394 sand-grained bioclastic fraction). They are always found isolated thus suggesting an
395 unpreserved substrate of possible phytal origin. The alveolinid *Flosculinella* (Fig. 13C) is present
396 in a few samples. *Gypsina* (Fig. 13E) is of common occurrence as well as other robust hyaline
397 foraminifers such as amphisteginids and miogypsinids. Based on the nature of epiphytic
398 (*Austrotrillina*, soritids, peneroplids and calcarinids) and potentially epiphytic (victoriellids)
399 taxa which largely dominate the foraminiferal assemblage, three foraminiferal associations have
400 been defined (Figs.12 and 14): 1) a porcelaneous-dominated epiphytic foraminiferal association
401 dominated by *Austrotrillina*, soritids and peneroplids (77 % in average of the epiphytic
402 foraminiferal composition, 45 % of the whole foraminiferal content) (Fig. 13A, B, C, F), 2) a
403 rotalid-dominated epiphytic foraminiferal association dominated by *Neorotalia* (88 % in average
404 of the epiphytic foraminiferal composition, 57 % of the whole foraminiferal content) (Fig. 13D)
405 and 3) a victoriellid-dominated association (60 % of the whole foraminiferal content) (Fig. 13E).
406 The coralline algal assemblage is composed of *Lithophyllum* gr. *pustulatum*, "*Lithophyllum*
407 *pseudoamphiroa*", *Corallina* and *Jania*. The green algae *Halimeda* is rare and has been observed
408 in a very few samples (Fig. 13B). Sand-grained to gravel-grained lithoclasts (altered peridotites)
409 mixed within the carbonate fraction are of common occurrence (Fig. 13B).

410 Scleractinian floatstones (FA6-2) are commonly interbedded within foraminiferal-coralline algal
411 packstone-grainstones and conglomerate thin (0.10–1.0 meter-thick) beds (Fig. 11D). The
412 foraminiferal assemblage is similar to that of the foraminiferal-coralline algal packstone-
413 grainstones (Table 1) since it is dominated by benthic foraminifers with a large proportion of

414 epiphytes and coralline algae including both articulated and encrusting forms. The benthic
415 foraminiferal association is composed of soritids, alveolinids (*Flosculinella*), miliolids
416 (*Austrotrillina*), calcarinids (*Neorotalia*) together with significant proportions of victoriellids.
417 amphisteginids, calcarinids and miogypsinids (*Miogypsina* and *Miogypsinoidea*) are present in
418 minor proportions. The scleractinian assemblage includes *Heliastrea*, *Stylophora*, *Goniopora*,
419 *Porites*, *Siderastrea* and *Dipastraea* (Maurizot et al., 2016). Corals are mainly found as rubbles
420 but rare *in-situ*, small-sized (< 50 cm thick, < 1m in diameter) coral bioconstructions (=
421 scleractinian boundstone: FA6-3) have been observed interbedded within coral floatstone
422 intervals (Fig. 11G).

423 A few kilometers away from the Wharf outcrops, at the S2 well, the B3 Member is significantly
424 thicker (80 m), strongly dolomitized (Fig 11E), and displays karst features filled by reddish clay-
425 rich sediments (Fig 11F).

426
427 Interpretation: in both foraminiferal-coralline algal packstone-grainstones and coral floatstones,
428 the foraminiferal assemblage is dominated by taxa which are known as common epiphytes (flat,
429 disk-shaped soritids, *Austrotrillina*, *Neorotalia*) and which suggests very shallow, euphotic
430 environments with extensive development of seagrass meadows (e.g. Chaproniere, 1975; Pomar
431 et al., 2014). Robust and rounded specimens of amphisteginids are also indicative of very
432 shallow-water settings (Hallock & Glenn, 1986) and their diameter-thickness ratios (<2) suggest
433 water-depth lower than 10 meters (Renema, 2005; Mateu-Vicens et al., 2009). Even though
434 calcarinids such as *Neorotalia* may live on sandy substrates, they have been shown to prefer
435 phytal substrate such as macro-algae (Renema, 2010; Reich et al., 2015), short filamentous algae
436 (Renema, 2003) and seagrasses (Lobegeier, 2002). *Neorotalia* has been shown to be abundant in
437 Oligo-Miocene seagrass facies (Fournier et al., 2004; Pomar et al., 2014). The *Flosculinella*-
438 *Austrotrillina*-soritid association has been reported in various Neogene seagrass meadows
439 environments (Chaproniere, 1975; Fournier et al., 2004; Hallock et al., 2006; Reich et al., 2014).
440 Additionally, the general absence of preserved substrate suggests that victoriellids encrusted

441 non-calcified biota such as, possibly, seagrass leaves or rhizomes. The relatively high-mud
442 content (packstone texture: Fig. 13A, B, C, D, E) also supports the interpretation of sea-grass
443 environments (e.g. Brandano et al., 2009; Reuter et al., 2012; Mateu-Vicens et al., 2012).
444 Additionally, abundance of articulated coralline algae is a common feature of seagrass meadow
445 environments (e.g. Lukasik et al., 2000; James & Bone, 2007) even though they may be common
446 in a wide range of shallow marine environments. Similar bioclastic composition present in
447 grainstones which may reflect a deposition in high-energy, non- or poorly-vegetated areas, at
448 vicinity of seagrass meadows.

449
450 The foraminiferal assemblage of the scleractinian floatstones suggests that coral fragments
451 probably came from nearby inner-shelf patch reef frameworks or could represent sparse coral
452 colonies within seagrass meadows. The few *in situ* coral bioconstructions observed in the B3
453 Népü Limestone Member suggest that they were probably relatively sparse on sea bottom of
454 relatively small dimensions (meter-scale).

455
456 As a consequence, the biotic composition and the textural characters of the FA6 facies
457 association suggest a deposition within the euphotic zone, in a relatively high energy
458 environment which is extensively vegetated by seagrass meadows and colonized by sparse and
459 small-sized coral bioconstructions. Additionally, the sub-rounded pebble lags which are
460 interbedded within carbonate deposits are an additional evidence for the proximity of the
461 ultramafic source in this mixed system and likely represent the reworking of fluvial outputs by
462 waves and marine currents (Nemec & Steel, 1984).

463

464 ***FA7- Subaqueous distributary channel fills (Fig. 15)***

465 Description: in the proximal part of the B3 Népü Limestone Member, numerous conglomeratic
466 channel-fills occur. They are particularly well-exposed on the Pindai Peninsula, at the

467 southeastern extremity of the Wharf outcrops (Fig. 15A), and to the South of the *Baie des Sapins*
468 (Fig. 15C). They cut into shallow water limestones and are 2 to 5 meters-thick and 15 to 30
469 meters-wide. At the Wharf outcrops, they are stacked into amalgamated channel complexes (Fig.
470 15A). Individual channel fills comprise well rounded, ultramafic pebbles-cobbles in a calcareous
471 matrix and coarse-grained gravelly sandstones (Fig. 15B). The composite infill is organized into
472 a fining upward, laterally accreting stack of inclined bedsets.

473

474 Interpretation: the interfingering of conglomerate and carbonate ramp deposits, the sharp
475 nature of erosional surfaces and the normally graded infills strongly suggest that these
476 conglomerates deposited in subaqueous distributary channels which formed in an alluvial-fan
477 front setting, by erosional currents transporting material from the Kopéto and Boulinda massifs
478 into a shallow-water carbonate inner ramp (e.g. Nemec & Steel, 1984). The lack of bidirectional
479 current structure would rule out a tidal origin for such channels. Such channels can be regarded
480 as the main feeders of the coarse terrigenous fraction present within the FA6 carbonate deposits
481 (conglomeratic lags).

482

483 ***FA8- Carbonate-dominated tidal flat (Fig. 16)***

484 Description: at the Wharf outcrops and the Grimault Islet (see location Fig. 4), the base of the B3
485 Népü Limestone Member is marked by an alternation of marls and centimeter-thick beds of very
486 fine-grained sandy to silty limestones with plant debris (Fig. 16A). Some of these silty marls
487 intervals are entirely cut by vertical pneumatophore roots (Fig. 16C). Among such alternations,
488 lithoclastic wackestone to packstone beds with fragments of echinoids and gastropods are
489 observed. In addition, several thin layers with decimeter-scale well rounded and well sorted
490 ultramafic pebbles in a calcareous muddy matrix are also identified (Fig. 16D). Finally, sandy
491 intervals display large scale sigmoidal dunes prograding towards the SSW to SSE. Foresets have
492 a 5° to 15° dip and bedforms extend over at least several tens of meters. These deposits consist

493 in medium-grained sandstones with floating granules, common bioturbation and erosional
494 topsets (Fig. 16B).

495

496 *Interpretation:* the alternation between marls and fine-grained sandstones with vegetal debris
497 and pneumatophore roots is seen as characteristic of a supratidal to intertidal environment
498 (Shinn, 1983). Pneumatophore roots are highly likely to evidence a supratidal swamp in a
499 tropical climate. Sigmoidal dunes are consistent with this interpretation and root traces and leaf
500 fragments suggest that they formed in a supra to intertidal environment such as tidal sand bars.
501 Conglomerates, in turn, would correspond to the terminal distributary channels of the alluvial-
502 fan system.

503

504

4.2 Vertical distribution of facies association in the S2 cores.

505
506 The contact between the Aquitanian Lower Népoui Formation and Burdigalian Upper Népoui
507 Formation is not observable at Chapeau Chinois outcrop (Fig.7 in Maurizot et al., 2016), but the
508 transition is continuously recorded at well S2 (Fig. 17A). This transition, between –104 m and
509 –98 m, shows the top of Lower Népoui Formation (Aquitanian after Maurizot et al., (2016)),
510 which is characterized by scleratinian flostones with a foraminiferal-algal packstone matrix.
511 The uppermost Lower Népoui limestones display in that location a rich foraminiferal
512 assemblage characterized by a regular occurrence of *Miogypsina*, *Heterostegina*, *Lepidocyclina*,
513 *Sorites*, *Neorotalia*, *Cibicides*, *Elphidium* and discorbids (Fig. 17B). The mixture of epiphytic
514 foraminifers such as *Sorites*, *Neorotalia*, *Cibicides*, *Elphidium*, discorbids and larger foraminifers
515 such as *Miogypsina*, *Lepidocyclina* and *Heterostegina* within the underlying limestones are
516 indicative of a euphotic to upper mesophotic depositional environment for the uppermost Lower
517 Népoui Formation (Pomar et al., 2014).

518 Between –98 m to –92 m depth, the transition from the Lower Népoui limestones to the
519 Upper Népoui basal sandstones is gradational, with a progressive increase in terrigenous clasts
520 and a decrease in biogenic components (Fig. 17C). At the top of this interval, fine to medium-
521 grained clean sandstones (FA4-FA5 facies associations) display parallel to cross laminations
522 with common bioturbation. A coarsening upward trend is evidenced between –92 m to –83 m,
523 with coarse- to medium-grained cross-bedded sandstones at the top. Constitutive elements
524 primarily comprise weathered serpentinite grains as well ferricrete clasts. Additionally, the
525 sandstones from the entire terrigenous interval (–98 m - –83 m) are strongly bioturbated and
526 contain a low-diversified foraminiferal assemblage (mainly broken and likely transported
527 *Miogypsina* and *Sorites*). The clean nature of the sand, the occurrence of cross-bedded beds and
528 the associated ichnofacies (*Thalassinoides-Dactyloidites*) suggest a deposition in a shallow,
529 subtidal, high-energy environment.

530 This terrigenous interval is overlain by the 80 meters-thick Népü Limestone Member
531 (B3), that is partially karstified and pervasively dolomitized (Fig. 17D). This interval displays

532 karst features infilled with reddish clay-rich sediment. Such reddish infills likely correspond to
533 reworked laterites of the post-Burdigalian Muéo Formation that unconformably overlay the
534 Miocene series (Coudray, 1975; Paris, 1981; Sevin et al., 2014).

535 The significant dolomitization of this interval could be explained by the high Mg content
536 of runoff water derived from weathered peridotite massifs in an area where mixing fresh and
537 marine water are already prone to induce secondary dolomitization (Tucker and Wright, 1990).

538
539 Finally, both the uppermost Lower Népoui and lowermost Upper Népoui formation were
540 deposited in shallow waters, within the euphotic to upper mesophotic environments. As a
541 consequence the upward transition from a carbonate-dominated to a siliciclastic-dominated
542 system is not accompanied by a major shift in paleowater depth.

543

544

545 **5. DISCUSSION**

546 **5.1 Revised depositional model for the Burdigalian Upper Népoui** 547 **mixed system**

548 This study proposes a revised depositional model of the Burdigalian Upper Népoui formation.
549 This constrained by seven synthetic logs, from S2 core data and 6 key outcrop sections (Fig. 18),
550 which capture the architecture of this mixed carbonate-siliciclastic system and its evolution
551 through time (Fig. 19).

552 Facies analysis and correlation of synthetic cross-sections suggest that alluvial fan and fluvial
553 deposits (FA1, FA2 and FA3) evolve distally into tidally influenced nearshore sandstone deposits
554 (FA4 and FA5) in only a few kilometers, in the lower part of the Upper Népoui formation (Fig.
555 18). Alluvial deposits are of restricted extension and the relatively sharp transition to marine
556 deposits together with the lack of preserved floodplain deposits strongly suggest a fan delta
557 depositional system (*sensu* Holmes, 1965). The good preservation of tidal deposits could also be

558 consistent with sedimentation occurring within a paleovalley (estuarine environment) or a
559 lagoonal embayment (Dalrymple et al., 1992). However, the lateral continuity of facies
560 associations along a 10 km-long transect (perpendicularly to a proximal-distal axis), between
561 the Pindai peninsula and the Beco Island (Fig. 18), favors a fan delta system developing on an
562 unconfined continental shelf. During this stage, continental alluvial fan deposits (FA1, FA2 and
563 FA3) would evolve laterally into a relatively wide and low-angle shelf (up to 5 km) where
564 intertidal (FA4) and subtidal (FA5) terrigenous sedimentation occurs (Fig. 19A). In the S2 well, a
565 relatively thin (<20m) siliciclastic unit, composed of facies associations FA5 and FA4
566 progressively changes upward into carbonate ramp deposits (FA6). This transitional facies
567 change suggests coeval terrigenous inputs and carbonate ramp sedimentation. Lateral transition
568 between shallow-water carbonates and siliciclastic deposits is observed in the modern lagoon of
569 New Caledonia where high fluvial discharges from peridotite massif watersheds (Garcin et al.,
570 2013) are coeval with high carbonate productivity (Chevillon, 1992; Cabioch et al., 2008; Le Roy
571 et al., 2019). Carbonate sedimentation in the B3 member is dominated by seagrass-related
572 carbonate production (FA6), suggesting that the subtidal, euphotic shelf was extensively covered
573 by seagrass meadows. In Pindai and Grimault sections the upward vertical transition from
574 terrigenous-dominated to carbonate-dominated sedimentation is characterized by the
575 development of a carbonate-dominated unit deposited in tidal-flat and swamp environments
576 (FA8). This suggests that tidal currents were able to import toward coastal environments (tidal
577 flat and swamps) significant volume of carbonate sediment produced in subtidal seagrass
578 environments. The interfingering of conglomerate-filled channels (FA7) within FA6 and FA8
579 carbonate deposits indicate that the coarse terrigenous fraction of the sediment was supplied by
580 the alluvial fan through distributary channels cross-cutting the tidal flat and the subtidal
581 seagrass meadows (Fig.19B). Similar conglomerate-filled feeder channels developing at the
582 mouth of river courses and cross-cutting shallow-water mixed carbonate-siliciclastic deposits
583 have been evidenced at the front of a Tortonian fan delta from the Vera Basin, Spain (Braga et al.,
584 2001). These channels, which are not related to relative sea level falls and emersion of the shelf

585 (i.e. incised valley fills), have been shown to feed turbidite lobes in the basin. Similarly, in the
586 Burdigalian Upper Népoui formation, conglomerate-filled channels cross-cutting seagrass
587 deposits (FA8) are subaqueous and developed during an overall transgressive trend. As a
588 consequence, the vertical and lateral evolution of facies within the Burdigalian Upper Népoui
589 formation requires the definition of two distinct depositional models for the mixed carbonate-
590 siliciclastic system: 1) a siliciclastic-dominated system, during relative sea level lowstand and/or
591 early transgression, characterized by an extensive terrigenous intertidal and subtidal domain
592 and by moderate seagrass carbonate production (Fig. 19A), 2) a carbonate-dominated ramp
593 system, during later transgression and possibly relative sea level highstand, typified by
594 extensive seagrass meadow development over the entire subtidal, euphotic domain (Fig. 19B).
595 Distal equivalents in mesophotic to oligophotic environments are not observed in the current
596 dataset.

597

598 **5.2 Factors controlling the depositional architecture of the Upper**

599 **Népoui Formation**

600

601 The depositional environments recorded by the Burdigalian deposits of Népoui suggest
602 significant creation of accommodation space as well as *in situ* changes of environmental
603 conditions. During the Burdigalian, the onset of siliciclastic sediment deposition (B1 Pindaï
604 Conglomerates and B2 Wharf Members) overlying the Aquitanian ramp is recorded in the S2
605 well by a progressive shift towards sandy terrigenous deposits with an upward decrease in
606 carbonate components (Fig. 17). This transition marks the demise of the Aquitanian carbonate
607 ramp, most likely due to a significant increase in water turbidity associated with the
608 progradation of alluvial fans. This facies shift has been formerly interpreted by Maurizot et al.
609 (2016) as resulting from a major forced regression driven by a tectonic uplift. However, the lack
610 of subaerial exposure evidence and the lack of significant paleobathymetric gap between the top
611 Aquitanian limestones (euphotic environment: 0–20m) and the base Burdigalian tidal sands in

612 S2 well do not support this interpretation of a forced regression at the Lower Népoui-Upper
613 Népoui transition. The increase of detrital fluxes and transient decrease or demise of carbonate
614 production during the Burdigalian in the Népoui area may result from other parameters such as:
615 1) an intensification of erosional and/or weathering processes on peridotite massifs due to
616 climate changes or to gravitational collapses linked with the fault network structuring Grande
617 Terre (Iseppi et al., 2018), or 2) an autocyclic lateral migration of a pre-existing alluvial fan
618 system. The overall transgressive trend recorded in the Pindai Peninsula by the upward
619 succession from fluvial to intertidal and finally subtidal carbonate ramp deposits suggests that
620 the Upper Népoui formation developed during a period of increasing accommodation space.
621 Published eustatic curves indicate an overall increase of eustatic sea level up to 80 meters
622 during the Burdigalian (Haq et al., 1987; Miller et al., 2005). Hence, eustatic variations may have
623 largely controlled the creation of accommodation space which allowed the accumulation of 60
624 meters (outcrops of Pindai Peninsula) to 100 meters (S2 well) of alluvial to shallow-water
625 (shoreface sands and euphotic seagrass carbonates) deposits. This creation of accommodation
626 space may also be linked to the collapse of the western margin driven by the post-obduction
627 isostatic rebound of Grande Terre (Lagabrielle & Chauvet, 2008, Collot et al., 2017).

628

629 **5.3 Seagrass and patch-reef carbonate production in high-terrigenous** 630 **influx settings**

631

632 A variety of carbonate-producing biota, including some large benthic foraminifera, coralline
633 algae, echinoderms, molluscs, and corals, tolerate nearly continuous siliciclastic influx (e.g.
634 Wilson, 2005; Novak et al., 2013; Santodomingo et al., 2015). The most distinctive character of
635 the lower Miocene carbonate sedimentation of the Népoui system is the dominance of seagrass
636 associated production, with also common occurrence of coral bioconstructions (meter-scale
637 patches). Seagrass-related carbonate production has been shown to be highly tolerant to high
638 terrigenous fluxes in various Cenozoic carbonate systems (e.g. Beavington-Penney et al., 2004;

639 Reuter et al., 2012; Brandano and Ronca, 2014; Tomas et al., 2016). In such settings, variations
640 in terrigenous supplies modify the taxonomic composition of plant assemblages on sea bottom
641 and therefore the morphology of phytal substrates for epiphytic foraminifers. In modern
642 environments, the nature and composition of epiphytic communities have been shown to
643 depend on the species of the seagrass substrate (e.g. Beavington-Penney et al., 2004; Perry et al.,
644 2012) or more commonly to the nature of the phytal substrate (seagrass vs macro-algae: Fujita
645 & Hallock, 1999 ; Benedetti & Frezza, 2016). Morphotypes of epiphytic foraminifers have been
646 revealed to be useful to analyze the assemblage structure and provide criteria to infer different
647 types of phytal substrates (Langer, 1993; Mateu-Vicens et al., 2010; 2014). Seagrasses are
648 typically dominated by flat, discoidal foraminiferal morphotypes (A and SB morphotypes after
649 Langer's (1993) classification, modified by Mateu-Vicens et al., 2014) since these foraminifers
650 have long life cycles (at least one year) and require long lasting and flat substrates. In contrast,
651 macroalgae, whose life span is shorter, are the substrate of smaller and robust foraminifers
652 (morphotypes B and D) with shorter life cycles (Mateu-Vicens et al., 2010). The porcelaneous-
653 dominated epiphytic association from Népoui is largely composed of SB morphotypes (flat and
654 discoidals soritids, peneroplids) and likely reflects a dominant seagrass substrate. In contrast,
655 the abundance of *Neorotalia* (morphotype B) in the calcarinid-dominated association would
656 characterize environments significantly colonized by macroalgae.

657 Repeated changes in epiphytic foraminiferal associations (porcelaneous-dominated vs calcarinid
658 (*Neorotalia*)-dominated) within the Népü limestone may therefore reflect changes through time
659 and space of the nature of the phytal substrate which in turn can reflect variations in some
660 environmental parameters such as turbidity or salinity. In modern tropical carbonate
661 environments, the dominance of calcarinids has been shown to be related to the extensive
662 development of macro-algae (Renema, 2010).

663 The Miocene seagrass skeletal association from Népoui is characterized by a high foraminiferal
664 (52 % of the sand-grained skeletal fraction in average) and coralline algal content (33 %), and an
665 extreme scarcity of *Halimeda* green algae. Such a bioclastic composition is consistent with the RA

666 (red-algal) foralgal seagrass assemblage (Brandano et al., 2019), which characterizes high
667 density tropical to subtropical seagrass meadows. RA-foralgal seagrass-related association is
668 reported from modern Mediterranean environments (Brandano et al., 2019) as well as from a
669 variety of ancient tropical carbonate systems (e.g. Seeb Formation, Eocene, Oman: Beavington-
670 Penney et al., 2004; Malampaya buildup, Philippines, Chattian to Burdigalian: Fournier et al.,
671 2004, SE Spain, Messinian: Sola et al., 2013). In densely vegetated areas, particularly in the case
672 of long-leaved seagrasses, oligophotic conditions prevail at sea bottom thus impeding the
673 development of green algae and promoting in contrast coralline algae, which are known to be
674 highly tolerant to low-light, oligo-mesotrophic conditions. Such an interpretation may explain
675 the high coralline content and the scarcity of *Halimeda* of seagrass sediments in the Népoui
676 seagrass-related sediment, but the oligophotic conditions at sea bottom can also result of
677 relatively high-turbidity created by strong terrigenous discharges. As a consequence it can be
678 suggested that in high-terrigenous settings, RA-foralgal associations may also occur in low to
679 moderately dense seagrasses as a result of water turbidity.

680 Seagrass meadows have a significant effect on current velocity reduction, sediment trapping and
681 decreasing resuspension of particles thus leading to a reduction of the turbidity in the water
682 column (de Boer, 2007). The reduction of fine particle resuspension within seagrass meadows
683 may have drastically limited the negative effect of water turbidity for the development of light-
684 dependent biota such as symbiont-bearing foraminifers, coralline algae and corals.

685 Differential trapping of sediments within the seagrass meadow may lead to slightly raised sea
686 bottom in vegetated areas (e.g. Pérès & Picard, 1964; Madsen et al., 2001; Fujita et al. 2015).

687 Water flow and terrigenous influx are therefore likely to be directed toward the lower lying
688 areas, thus reducing the risk of suffocation for filter-feeding biota living in the seagrass meadow.

689 The coeval development of seagrass meadows and terrigenous sediment flows on the alluvial-
690 fan front possibly also controls the aggrading nature (vertical stack) of the main distributary
691 channel infills (Fig 15).

692 Additionally, scleractinian bioconstructions that form in turbid water under significant
693 terrigenous inputs, are common in the geological record (e.g. Sanders & Baron-Szabo, 2005) and
694 particularly in Cenozoic shallow-water sedimentary systems from the Indo-Pacific (e.g. Wilson,
695 2005; Novak et al., 2013; Santodomingo et al., 2015) and Mediterranean regions (Galloni et al.,
696 2001; Brandano et al., 2010, 2016; Mazzucchi & Tomassetti, 2011). This is consistent with the
697 common occurrence of coral fragments and in situ bioconstructions within the Lower and Upper
698 Népoui limestones. As evidenced from Miocene patch-reefs formed at the seaward margin of the
699 equatorial Mahakam Delta (Wilson, 2005), carbonate production and patch-reef development
700 may occur in turbid-water, delta-front areas during any phase of eustatic sea level.

701 The relative tolerance of seagrass ecosystems as well as small-scale coral patches environments
702 to high terrigenous influx may explain the significant development of ramp carbonate systems
703 with mixed foraminiferal and scleractinian production at the mouth of alluvial fan systems, in
704 tropical settings.

705

706

707 **6. CONCLUSION**

708

709 The Burdigalian carbonate ramp system of Népoui is a well-constrained example of seagrass-
710 dominated and coral-bearing carbonate systems developing in high terrigenous influx setting
711 and tropical climate.

712 (1) During the Burdigalian, the upward succession from the Pindaï conglomerate to the Népü
713 Limestone member is the result of the retrogradation or lateral migration of a fan delta system
714 passing laterally into a carbonate ramp, during a marine transgression. Despite this
715 transgressive trend, the carbonate ramp received high-terrigenous influx as evidenced by small-
716 scale conglomeratic channels that randomly incised the carbonate ramp.

717 (2) Post-obduction isostatic rebound resulted in an extensional tectonic regime and margin
718 collapse that possibly favored the onset and preservation of terrigenous fan delta and carbonate
719 ramp systems.

720 (3) Extensive development of seagrass meadows and potentially of macro-algae occurred in
721 shallow, euphotic environments subject to high-terrigenous discharge. Phytal substrates
722 induced by seagrass vs macro-algae colonization of sea bottom strongly controlled the nature of
723 carbonate production and promoted the accumulation of foraminiferal-coralline algal sediments.

724 (4) Finally, the development of seagrass meadows has probably largely contributed to the
725 preservation of a diverse and significant carbonate production in a context of strong terrigenous
726 inputs, by reducing water turbidity and by limiting the risk of suffocation for filter-feeding biota.

727 The lateral and vertical variability of the Burdigalian carbonate ramp improves knowledge on
728 the heterogeneity of potential hydrocarbon reservoirs in mixed carbonate-siliciclastic systems.

729

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1208 shelf profiles. *Sedimentology*, v. **62**, no. **7**, p. 2009–2037.
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1212 **FIGURE AND TABLE CAPTIONS**

1213 **Figure 1:** Simplified geological map of New Caledonia (modified after Maurizot & Vendé-Leclerc,
1214 2009) with location of the study area, western limit of obducted peridotites, drilling sites on the
1215 quaternary barrier reefs (Coudray, 1975; Cabioch et al., 2008; Montagianni et al., 2011), together
1216 with map extent of the Deep-Water Fold and Thrust Belt of the western margin of New
1217 Caledonia (DWFTB, Collot et al., 2017). See text for details of these features.

1218
1219 **Figure 2:** (A) Geological map of the Népoui are modified from Maurizot & Vendé-Leclerc (2009),
1220 (B) cross section through the Kopeto massif and the Népoui miocene outcrops. The Lower
1221 Miocene mixed carbonate-siliciclastic system overlies unconformably the Eocene Népoui Flysch
1222 and the Poya Terrane on the western Caledonian margin. The Peridotite Nappe structurally
1223 overlies the Poya Terrane and the Nepoui Flysch.

1224
1225 **Figure 3:** Synthetic sedimentary log combining observations from the S4 well and outcrop
1226 sections. The Lower Miocene mixed system is composed of the Lower Népoui Formation
1227 (Aquitanian) and the Upper Népoui Formation (Burdigalian) dated by biostratigraphic analysis
1228 and strontium isotope measurements (Maurizot et al., 2016). Both formations are divided into
1229 three members A1 to A3 and B1 to B3 characterizing respectively a carbonate ramp and a mixed
1230 inner ramp delta system.

1231
1232 **Figure 4:** Location map of dataset used in this study. See location on Fig 1 and 2.

1233
1234 **Figure 5:** Description, sedimentary processes and summary depositional environment
1235 interpretation of the different facies associations of the Upper Népoui Formation (Burdigalian).

1236
1237 **Figure 6:** Proximal alluvial fan (FA1)

1238 **A.** Massive, sandy matrix supported, poorly sorted conglomerate of the NE Pindaï
1239 Peninsula.

1240 **B.** Altered serpentinite pebbles with silicified cortex and greenish sands.

1241 **C.** Large block of dolerite that reach a diameter of 1 m.

1242 **D.** Gravelly sandstones with faint trough cross-bedding highlighted by gravels and
1243 pebbles.

1244 **E.** “Piste de roulage” outcrop (NW Muéo Peninsula) showing the unconformity between
1245 the Népoui Flysch (Priabonian) and the Upper Népoui Formation (Burdigalian) (see location on
1246 Fig.4).

1247 **F.** Miocene conglomerates of the Upper Népoui Formation organized in two normally
1248 graded sequences as shown on the sedimentological section.

1249

1250 **Figure 7:** Distal alluvial fan (FA2)

1251 **A.** Oblique bedded conglomerates that vertically evolve into trough cross-bedded
1252 sandstones capped by a planar abandonment surface.

1253 **B.** Meter scale oblique bedsets showing an apparent progradation towards the SSW.

1254 **C.** Red to greenish-red trough cross-bedded sandstones with root traces and iron oxide
1255 concretions

1256 **D.** Silicified wood pieces.

1257

1258 **Figure 8:** Fluvial channel fills (FA3)

1259 **A.** Interpretation of the depositional architecture of two erosional channels in the
1260 northern part of the Didot Islet (see location in Fig.4)

1261 **B.** Conglomeratic channel fill

1262 **C.** Conglomerates with large pebbles within a coarse to very coarse-grained sandy matrix
1263 organized in graded beds with erosional trough-shaped bases.

1264 **D.** Large silicified wood trunk.

1265

1266 **Figure 9:** Tidal sands (FA4)

1267 **A.** Small sigmoidal dunes showing tidal bundling and small scale trough-cross ripples

1268 between lamina sets

1269 **B.** Herring-bone cross-bedding

1270 **C.** Gravel and pebbles lags within a carbonate muddy matrix

1271 **D.** Greenish coarse to medium-grained cross-bedded sandstones.

1272 **E.** Thin section from the S2 well cores (85.6 m depth) showing altered serpentinite
1273 grains with radial fibrous cement and drusy cement.

1274

1275 **Figure 10:** Subtidal sands (FA5)

1276 **A.** Fine-grained bioturbated sandstones with planar laminations and small-scale ripples

1277 **B.** Isolated unidirectional ripples

1278 **C.** *Dactyloidites peniculus* burrows in fine-grained sandstones

1279 **D.** Fine-grained bioturbated and laminated sandstones.

1280

1281 **Figure 11:** Subtidal, seagrass related low-angle carbonate ramp (FA6)

1282 **A.** Outcrop of the southern part of Grimault Islet (see location Fig. 4).

1283 **B.** Focus on the southern part of Grimault Islet outcrop that shows foraminiferal-
1284 coralline algal packstone-grainstone facies.

1285 **C.** Calcitic tubes are sparsely distributed over carbonate ramp deposits. They could be
1286 interpreted as tube worms.

1287 **D.** Isolated corals and lags of ultramafic sub-angular pebbles and cobbles in a carbonate
1288 matrix.

1289 **E.** Thin section from the S2 well core (41.25 m depth) showing two types of dolomite
1290 crystals: clear large rhomboidal crystals and small rhomboidal crystals with cloudy centers.

1291 **F.** Karstified and dolomitized facies of the Nepü Member from S2 well cores (see location
 1292 on Fig.4). Karst features and fractures are filled by mustard yellow clay, carbonate fragments
 1293 and impregnated red mud.

1294 **G.** Small-sized *in situ* coral bioconstructions embedded within coral floatstones and
 1295 foraminiferal-coraline algal packstones-grainstones.

1296

1297 **Figure 12:** Simplified logs of the nine shallower drilled boreholes of the Pindai Peninsula and
 1298 Grimault Islet (see location Fig.4). Stars represent the locations of the thin sections.

1299

1300 **Figure 13:** Seagrass-related carbonate microfacies from the Burdigalian of Népoui (FA6) (see
 1301 location of thin sections on Figure 12):

1302 **A.** Foraminiferal-coraline algal packstone with abundant disk-shaped soritids (*Sor.*)
 1303 and miliolids including *Austrorillina* (*Austr.*). Grimault Island, well GR07, depth: 11.10
 1304 m.

1305 **B.** Foraminiferal-coraline algal packstone with disk-shaped soritids (*Sor.*), small
 1306 rotalids (*Rot.*), Halimeda (*Hal.*), articulated coralline algae (ARA), Planorbulina (*Plan.*)
 1307 and lithoclasts (*Lith.*). Pindai Peninsula, well 99N014, depth: 27.80m;

1308 **C.** Foraminiferal-coraline algal packstone with disk-shaped soritids (*Sor.*),
 1309 *Austrorillina* (*Austr.*), the alveolinid *Flosculinella* (*Alv.*), articulated coralline algae
 1310 (ARA) and Small miliolids (*Mil.*) Pindai Peninsula, well 96N004, depth: 1.50m;

1311 **D.** Foraminiferal-coraline algal packstone with small rotalids (*Neorotalia*) (*Rot.*),
 1312 articulated coralline algae (ARA) and lithoclasts (*Lith.*). Pindai Peninsula, well 99N014,
 1313 depth: 27.80m;

1314 **E.** Foraminiferal-coraline algal packstone with encrusting foraminifers Victoriellids
 1315 (*Vict.*), *Gypsina* (*Gyps.*) and small rotalids (*Rot.*). Pindai Peninsula, well 96N002, depth:
 1316 11.10m;

1317 **F.** Foraminiferal-coraline algal grainstone with disk-shaped soritids (*Sor.*), rotalids
 1318 (*Rot.*), the alveolinid *Flosculinella* (*Alv.*) and articulated coralline algae (ARA). Grimault
 1319 Island, well GR07.

1320

1321 **Figure 14:** Skeletal composition of the sand-grained skeletal fraction of seagrass-related
 1322 carbonates (FA6) using point counting on thin-sections.

1323

1324 **Figure 15:** Subaqueous distributary channel fills (FA7).

1325 **A.** The Wharf outcrop (see location Fig.4) showing a channel complex incising inner
1326 carbonate ramp and tidal flat deposits.

1327 **B.** Close-up on carbonate matrix-supported conglomerates that fill channels.

1328 **C.** The *Baie des Sapins* outcrop (see location Fig. 4) displays meter-scale isolated channels
1329 incising into carbonate ramp deposits.

1330

1331 **Figure 16:** Carbonate-dominated tidal flat deposits (FA8)

1332 **A.** Marls interbedded with very fine-grained sandstones with plant debris.

1333 **B.** Oblique sets of very bioturbated gravelly sandstones with truncated topsets.

1334 **C.** Marls interbedded with very fine-grained sandstones and lithoclastic limestones
1335 intersected by pneumatophores roots.

1336 **D.** Lag of well-sorted ultramafic pebbles in a carbonate matrix.

1337

1338 **Figure 17:** (A) Synthetic sedimentary log of S2 well (scale 1/500) (see location Fig.4). (B). Top
1339 of Lower Népoui Formation (Aquitanian), between -104 and -98 m, characterized by scleratinian
1340 flostones with a foraminiferal-algal grainstone-packstone matrix containing miogypsinids
1341 (*miog.*) and articulated coralline algae (*ARA*) (C). The lower part of the Upper Nepoui Formation
1342 (Burdigalian), between -98 to -83 m, characterized by fine to coarse-grained sandstones
1343 coarsening upward (FA4-FA5) that contain weathered serpentinite grains (Litho.), ferricrete
1344 clasts and few benthic foraminifers (*miog.*: miogypsinids). D. Népü Limestone Member (B3),
1345 between -80 to 0 m, showing pervasive karstification and laterite mud (*lm*) infills on cores. The
1346 thin section (-103 m) exhibit significant dolomitization features with few preserved bioclasts
1347 (RA: coralline algal fragments)

1348

1349

1350 **Figure 18:** Synthetic sedimentary sections (scale: 1/500) revealing the frontal and lateral facies
1351 variability of the Upper Népoui Formation (Burdigalian).

1352 **A.** Longitudinal profile within the proximal-distal sections of Muéo Peninsula, Pindaï-
1353 Wharf and Pindaï S2 core sections.

1354 **B.** Lateral profile across the Burdigalian mixed system with the Pindaï- Wharf, Grimault,
1355 Didot, Longue and Beco sections.

1356

1357 **Figure 19:** Depositional models for the mixed carbonate-siliciclastic system of Népoui
1358 (Burdigalian). (A) The burdigalian mixed system is first dominated by short alluvial fan deposits
1359 that laterally evolve distally into tidally influenced nearshore sandstone deposits and towards
1360 shallow water carbonate dominated by seagrass-related biota and corals derived from small-
1361 sized bioconstructions, and (B) is then overlain by seagrass-related carbonate ramp cut by
1362 several small-scale conglomerate channels.

1363

1364 **Table 1:** Skeletal composition of the sand-sized fraction (in %) of FA6 limestones (seagrass-
1365 related facies association) from point counting on thin-sections. The location of wells and
1366 outcrop samples is reported on Figure 4. Texture code: P=Packstone; G=Grainstone; FP:
1367 floatstone with packstone matrix. Foraminiferal assemblage code: p = porcelaneous-dominated
1368 epiphytic assemblage (*Austrotrillina*+ soritids + peneroplids); r = rotalid-dominated epiphytic
1369 foraminiferal assemblage; v = victoriellid-dominated foraminiferal assemblage.

1370

1371 **Table 2:** Skeletal composition of the sand fraction within FA6 limestones as a function of
1372 depositional texture and foraminiferal assemblage.

1373

Skeletal composition (%)

Sample	Dominant texture	Foraminiferal assemblage	Skeletal composition (%)															
			<i>Austrorillina</i>	Other miliolids	Soritids (<i>Sorites</i> , <i>Amphisorus</i>)	Peneroplids	Alveolinids	Agglutinated foram.	Victoriellids	<i>Neorotalia</i>	Other hyaline benthic foraminifers	Articulated coralline algae	Encrusting coralline algae	<i>Halimeda</i>	Mollusks	Echinoderm	Bryozoan	indeterminate
Well 96N002 : 4.00 m	P	p	6	27		8	1	12			4	7	19		7	2		7
Well 96N002 : 6.00 m	G	p	12	9	11			2			2	18	30			11	2	3
Well 96N002 : 8.10 m	G	p	4	17	21			7		9	9	1	27			1		4
Well 96N002 : 11.10 m	P	v						47	16	11	2	12			2	7		3
Well 96N002 : 19.80 m	P	r		12	5				33	13	10	21						6
Well 96N004 : 1.50 m	P	p	1	15	12	5	1	5		8	2	28	14		2	3		4
Well 96N004 : 3.50 m	P	r		3	10			3	3	17	1	37	14		1	4		7
Well 96N007 : 21.50 m	P	r		6				7	47	6	12	12			3			7
Well 99N008 : 24.70 m	P	r		2	2	1		4	41	9	9	14			5	6		7
Well 99N008 : 9.40 m	P	p	1	4	22			1	7	2	1	11	43		2	3		3
Well 99N008 : 15.00 m	P	p		3	13				10	3	4	14	14		9	26		4
Well 99N012 : 8.80 m	P	v						1	39	39		6	8		1	3		3
Well 99N014 : 18.50 m	P	r		12	8			5	1	21	4	14	17		4	6		8
Well 99N014 : 27.80 m	P	p		2	19			2	15	6	15	13	8		2	7	2	9
Well 99N015 : 16,25 m	FP	r		5				2	2	43	5	10	22		2	2		7
Well 99N017 : 11.75 m	P	p	6	5	21			7	2		6	21	15		2	6		9
Well 99N017 : 22.00 m	P	p		3	22			4	2	14	6	16	14		5	7		7
Well 99N017 : 22.05 m	P	p	4	6	18			7	4	9	6	12	18		2	6		8
Well GR07 : 11.10 m	G	p			25			2	4	3	1	29	25		2	4		5
CS26 (outcrop)	G	r	4	5	1			10	6	21	4	23	9		0	9		8
CS40 (outcrop)	FP	v		5	13		2	4	20	8	4	14	16		2	5		7
045_01 (outcrop)	G	p		9	22			2		4	8	7	33		2	6		7

Table 1: Skeletal composition of the sand-sized fraction (in %) of FA6 limestones (seagrass-related facies association) from point counting on thin-sections. The location of wells and outcrop samples is reported on Figure 4. Texture code: P=Packstone; G=Grainstone; FP: floatstone with packstone matrix. Foraminiferal assemblage code: p = porcelaneous-dominated epiphytic assemblage (*Austrorillina*+ soritids + peneroplids); r = rotalid-dominated epiphytic foraminiferal assemblage; v = victoriellid-dominated foraminiferal assemblage.

		Skeletal composition (%) of the sand-sized fraction									
		Epiphytic foraminifers*	Victoriellids	other benthic foraminifers	articulated RA	encrusting RA	Halimeda	Mollusk	Echinoderm	Bryozoan	indeterminate
All samples N=22	Average	30	7	15	14	19	1	3	6	0	6
	min	14	0	1	1	9	0	0	0	0	3
	max	47	47	33	37	43	14	9	26	2	9
Grainstones N=5	Average	27	2	17	16	25	0	1	6	0	5
	min	23	0	3	1	9	0	0	1	0	3
	max	34	6	33	29	33	0	2	11	2	8
Packstones N=17	Average	30	8	14	13	17	1	3	5	0	6
	min	14	0	1	2	12	0	0	0	0	3
	max	47	47	32	37	43	14	9	26	2	9
Samples with porcelaneous- dominated epiphytic foraminiferal assemblage (N=12)	Average	27	3	16	14	22	2	3	7	0	6
	min	14	0	3	1	13	0	0	1	0	3
	max	36	12	33	29	43	14	9	26	2	9
Samples with calcarinid-dominated epiphytic foraminiferal assemblage (N=7)	Average	36	3	16	16	16	0	2	4	0	7
	min	26	0	7	9	9	0	0	0	0	6
	max	47	7	25	37	22	0	5	9	0	8
Samples with victoriellid-dominated foraminiferal assemblage (N=3)	Average	24	34	9	7	14	0	2	5	0	4
	min	16	20	1	2	12	0	1	3	0	3
	max	36	47	15	14	16	0	2	7	0	7

Table 2: Skeletal composition of the sand fraction within FA6 limestones as a function of depositional texture and foraminiferal assemblage.

Figure 1

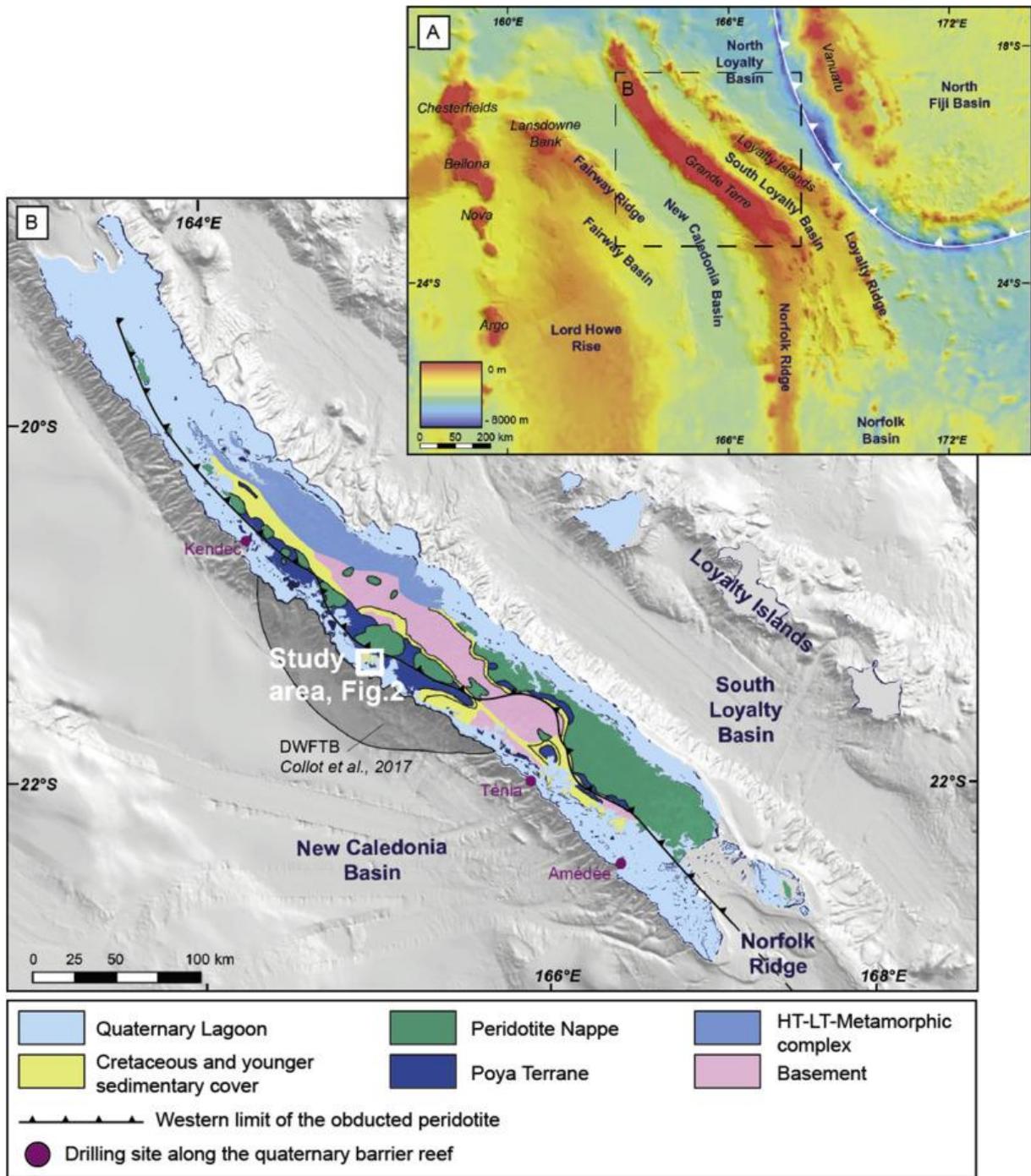


Figure 2

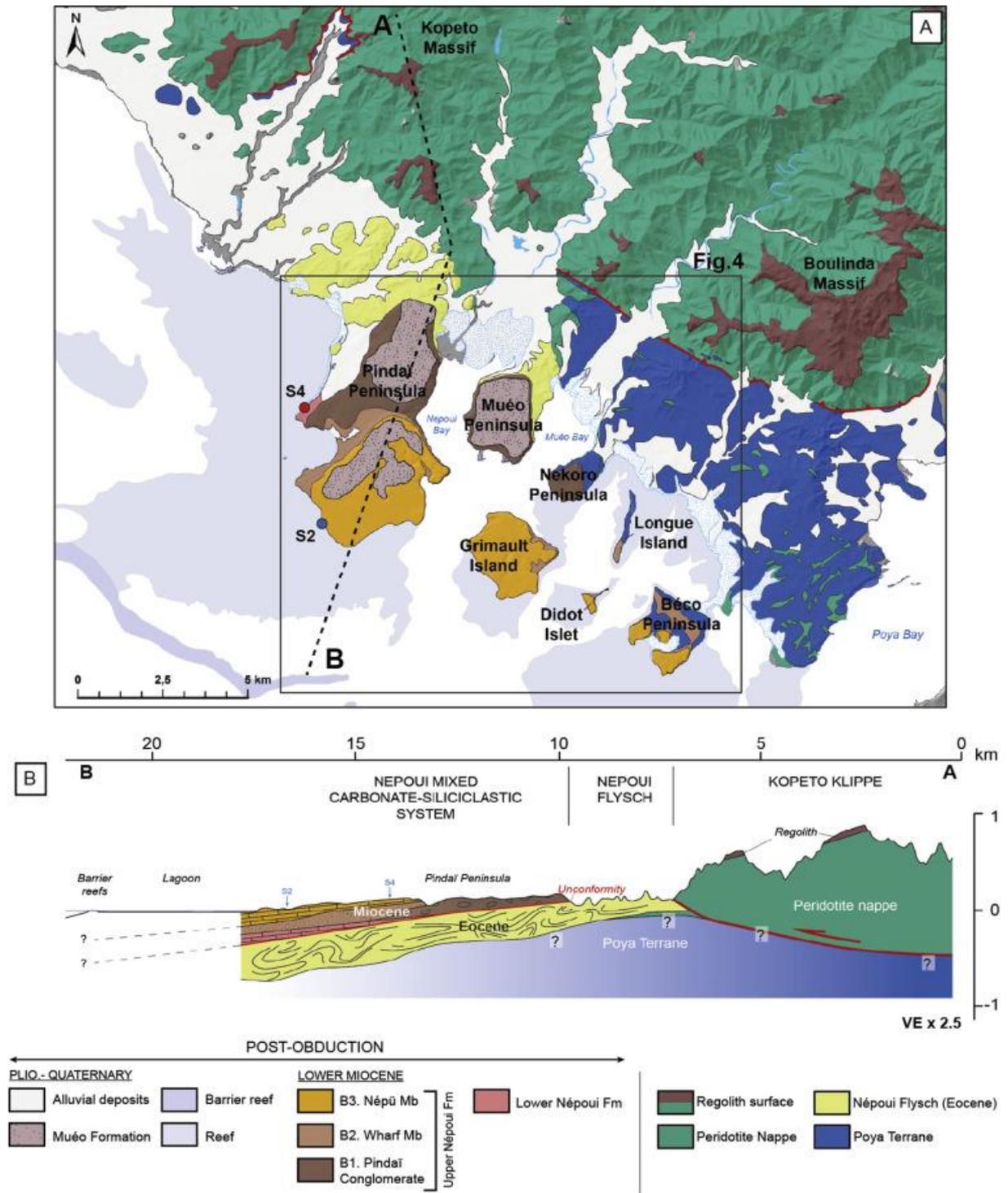


Figure 3

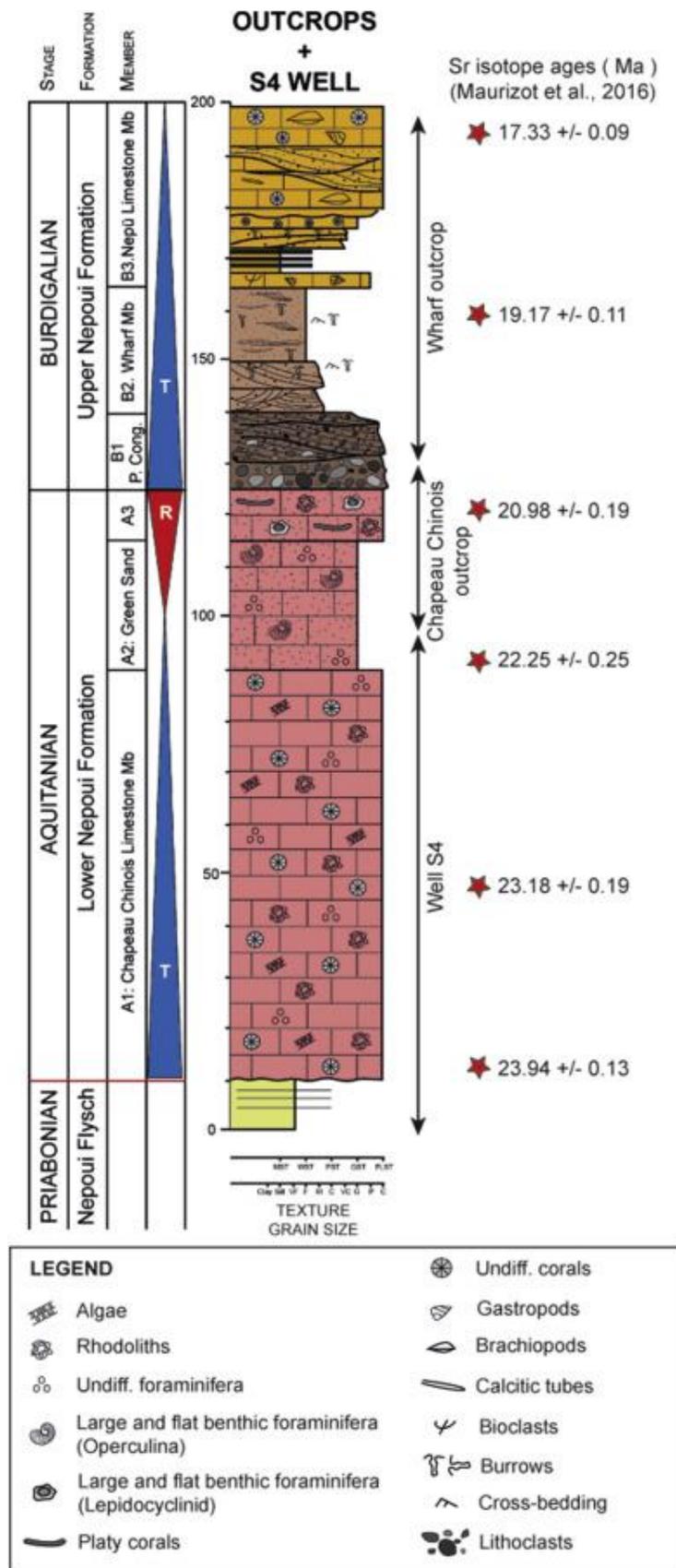


Figure 4

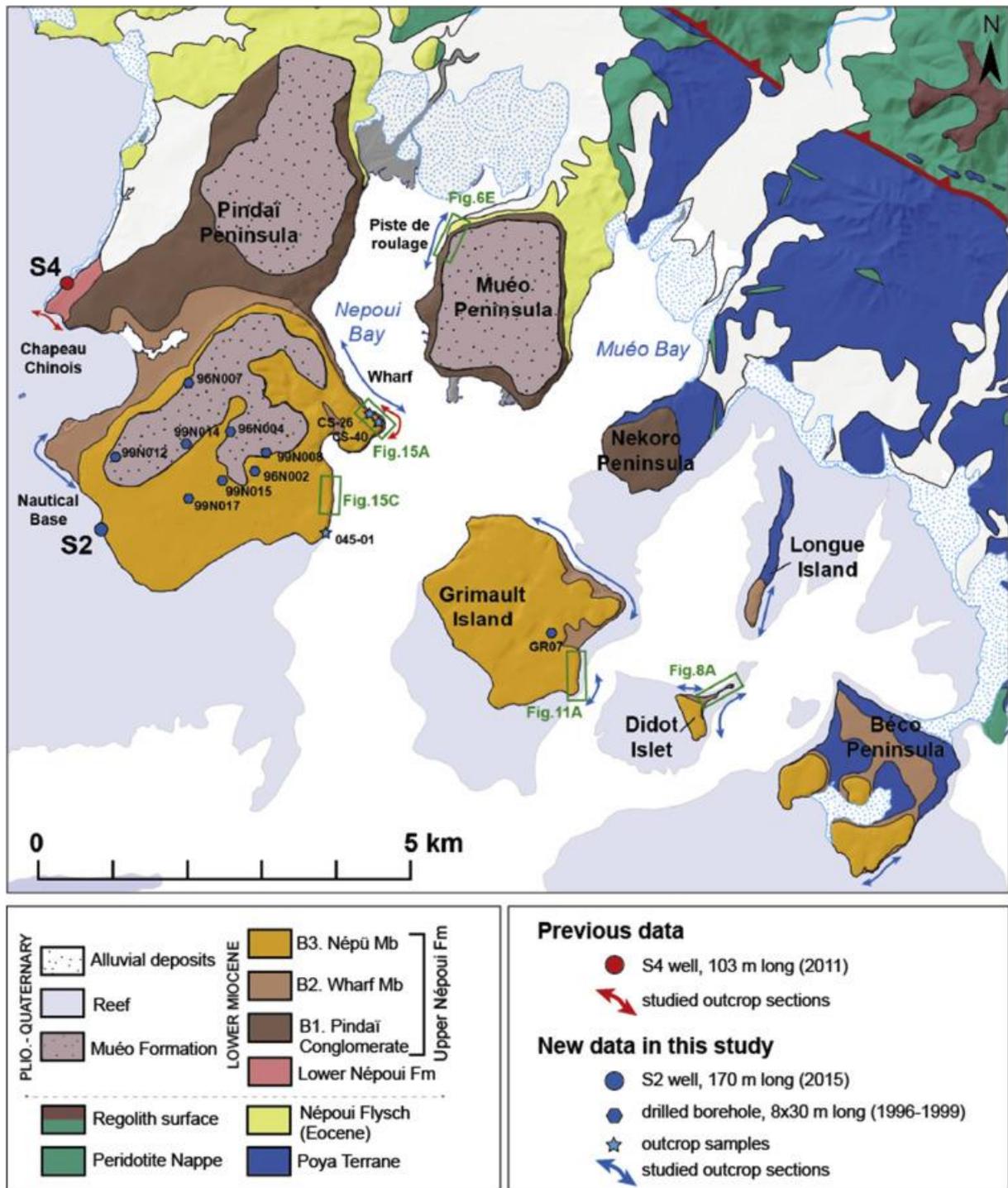


Figure 5

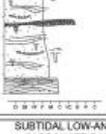
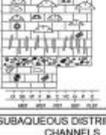
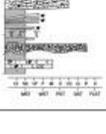
Mb.	Facies association	Description	Interpretation	CONSTITUENTS
B1. Pindar Conglomerate Mb.	FA-1 PROXIMAL ALLUVIAL FAN 	Disorganized pebble-boulder conglomerates and normally graded conglomerates that vertically evolve toward trough cross-bedded gravelly sandstones.	These successions are interpreted as subaerial debris-flow deposits of alluvial fan (Blair & McPherson 1994; Blair 1999) that evolve toward bedload deposits within a coarse braided river system that forms barforms or channel fills (Bluck, 1979; Miall, 1985).	<p>CONSTITUENTS</p> <ul style="list-style-type: none"> Fe concretions Burrows Roots, root traces Woods Plant debris Current ripples Large coral Undiff. corals Gastropods Bivalves Urchins Echinoderms Calclitic tubes Red algae Rhodoliths Red laminae Porcelaneous benthic foraminifers <p>GRAIN SIZE</p> <ul style="list-style-type: none"> Clay Silt Very Fine (V.F) Fine to Medium Coarse to Very Coarse (V.C) Gravel (2 to 4 mm) <p>Conglomerates</p> <ul style="list-style-type: none"> 40 to 100 mm 100 to 300 mm 300 to 500 mm <p>TEXTURE</p> <ul style="list-style-type: none"> WST: wackestone PST: packstone GST: grainstone FLST: floatstone
	FA-2 DISTAL ALLUVIAL FAN 	Oblique laminated conglomerates with meter-scale oblique and sigmoid sets with a SSW direction that evolve in through cross-bedded sandstones with pieces of wood, root traces and iron oxide concretions.	These normally graded sequences could result from the sedimentary discharge of high capacity flows that would form sand-gravel bars and three dimensional dunes (Allen, 1983; Miall, 1985) as frontal splays of alluvial fans.	
	FA-3 FLUVIAL CHANNEL FILLS 	Channels filled by cross-stratified conglomerates with large pieces of wood. The channels are overlain by very fine sandstones interbedded with marls.	These conglomerates are interpreted as channelised bedload deposits within coarse braided river system (Allen, 1963; Miall, 1977). These channels are probably located near the coastline in the fluvial-marine transition domain.	
B2. Wharf Mb.	FA-4 TIDAL SANDS 	Medium to very coarse-grained trough cross-bedded sandstones with erosional surfaces and few mud drapes, mud clasts and bioturbations interbedded with lags of pebbles within a carbonate matrix showing common loading features. Cross-bedded sandstones can display clear sigmoidal tidal bundles.	These cross-bedded sandstones could be interpreted as tidal sand bars within tide influenced delta front deposits (Holmes, 1965) or an estuarine environment (Belderson et al., 1982; Reynaud and Dalrymple, 2012). Conglomeratic lags are interpreted as episodic fluvial inputs into shallow marine environments, possibly during floods (Mutti et al., 2000).	
	FA-5 SUBTIDAL SANDS 	Fine to very fine-grained tabular cross-bedded sandstones with bioturbations, ripples and lags of gravels. Where grain size is finer cross bedding is replaced by parallel laminations and isolated unidirectional ripples.	The bioturbated sandstones with oblique sets and ripples could be interpreted as subtidal sands deposited at river mouths (McPherson et al., 1987). Sandy materials could be transported by fluvial or alluvial fan processes into the sea. The strong vertical and horizontal bioturbation intensity suggest relatively quiet conditions in the marginal part of the fan delta front.	
B3. Népu Mb.	FA-6 SUBTIDAL LOW-ANGLE CARBONATE RAMP 	Porcelaneous foraminifera coralline algal packstone-grainstone and floatstone with a coralline algal-porcelaneous foraminiferal packstone that are affected by dolomitisation and karstification processes in the distal part of the Nepou system. Limestones are interbedded with several lags of ultramafic sub-angular pebbles-cobbles comprised within a carbonate matrix.	This limestone facies association is interpreted as being deposited in shallow euphotic inner ramp environments associated with seagrass meadows with sparse coral colonies (Maurizot et al., 2016). Ultramafic pebble-cobble lags could correspond to the reworking of fluvial outputs by waves and marine currents.	
	FA-7 SUBAQUEOUS DISTRIBUTARY CHANNELS 	Channels filled by normally graded, imbricated pebble-cobble conglomerates within a carbonate muddy matrix typically organised as laterally migrating sets. Channels incise carbonate ramp limestones.	Conglomerates deposited in subaqueous terminal distributary channels of a fan-delta which, incised a shallow-water carbonate inner ramp.	
	FA-8 CARBONATE-DOMINATED TIDAL FLAT 	Marls interbedded with very fine sandstones with plant debris and pneumatophore roots and some lithoclastic wackestone-packstone limestone beds with mollusks and benthic foraminifera. Several ultramafic clastic sediments occur and form lags of polygenic well-sorted pebble conglomerates, and sigmoid oblique laminated strongly bioturbated gravelly sandstones.	These sequences are characteristic of supratidal to intertidal environments (Shinn, 1983). The numerous pneumatophore traces point to deposition within a mangrove environment in a supratidal domain whereas lithoclastic limestones with mollusks and benthic foraminifera would suggest deposition within an intertidal domain. Finally, the sigmoid oblique laminated gravelly sandstones could be formed in the inner carbonate ramp during flood discharge events.	

Figure 6

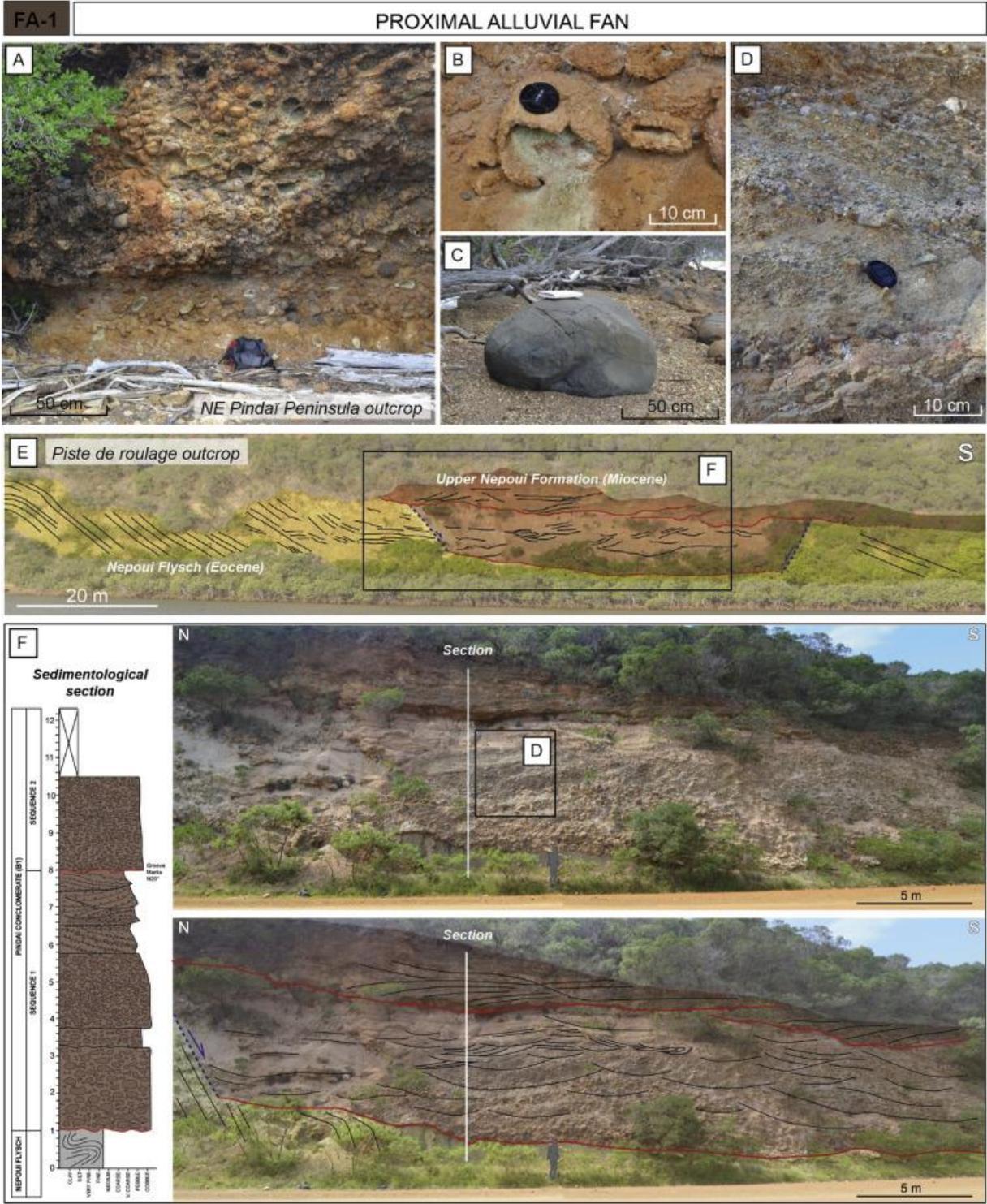


Figure 7

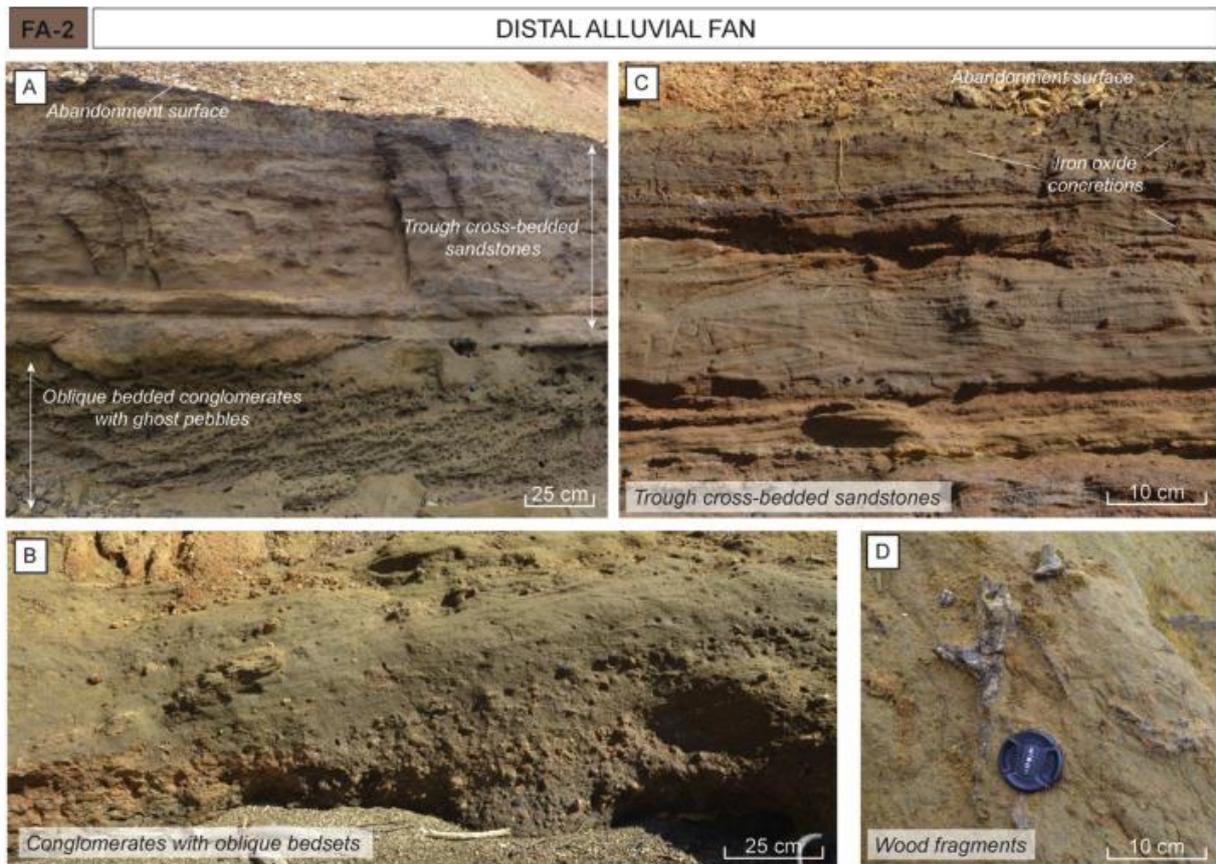


Figure 8

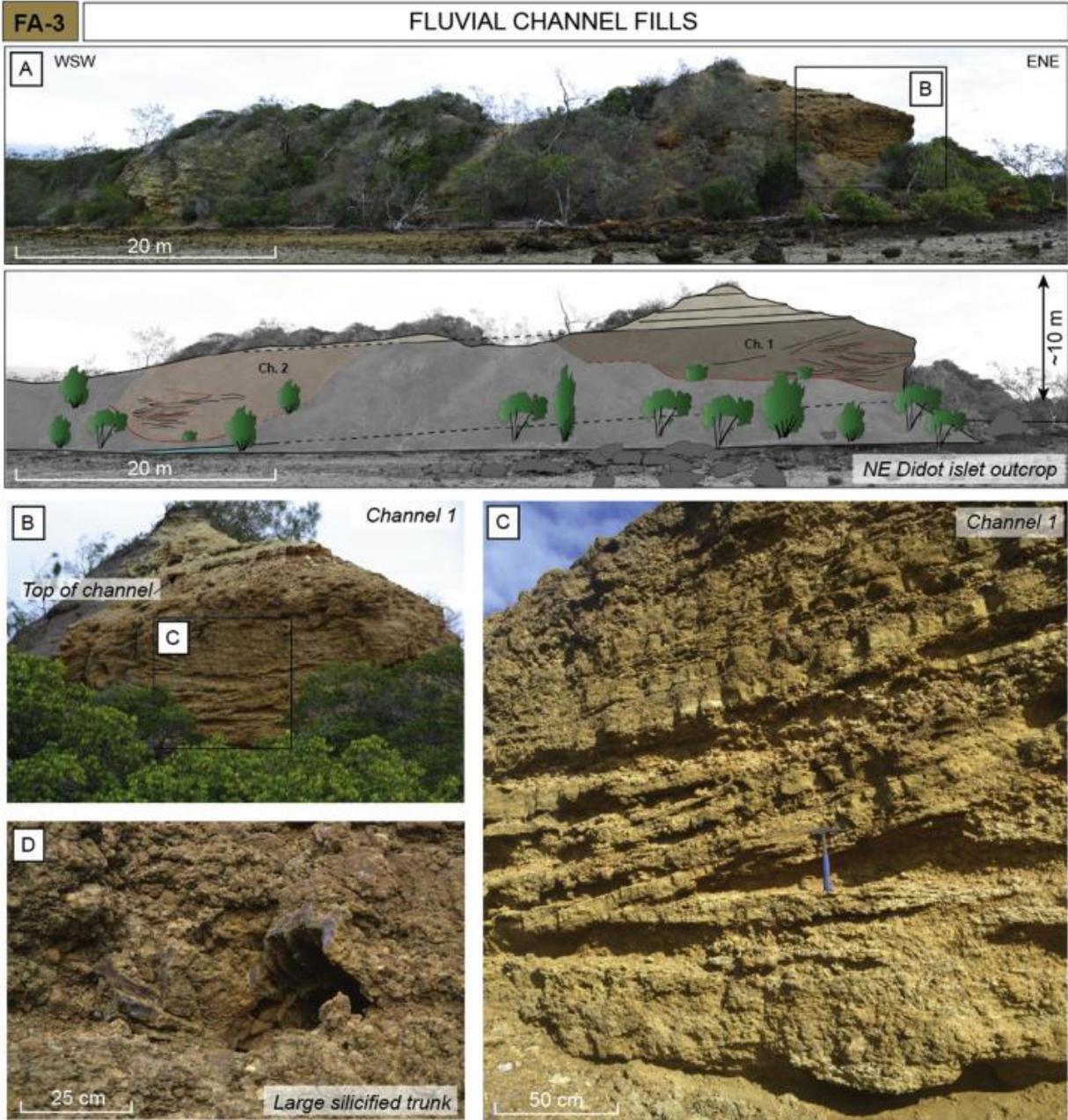


Figure 9

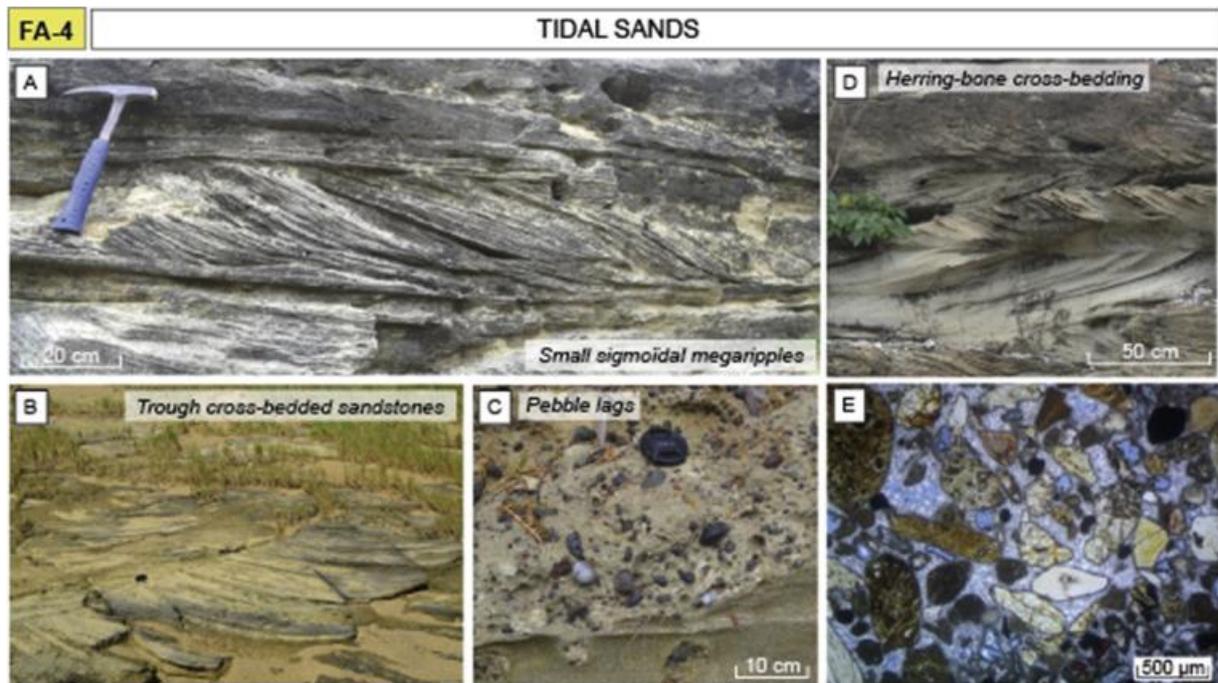


Figure 10

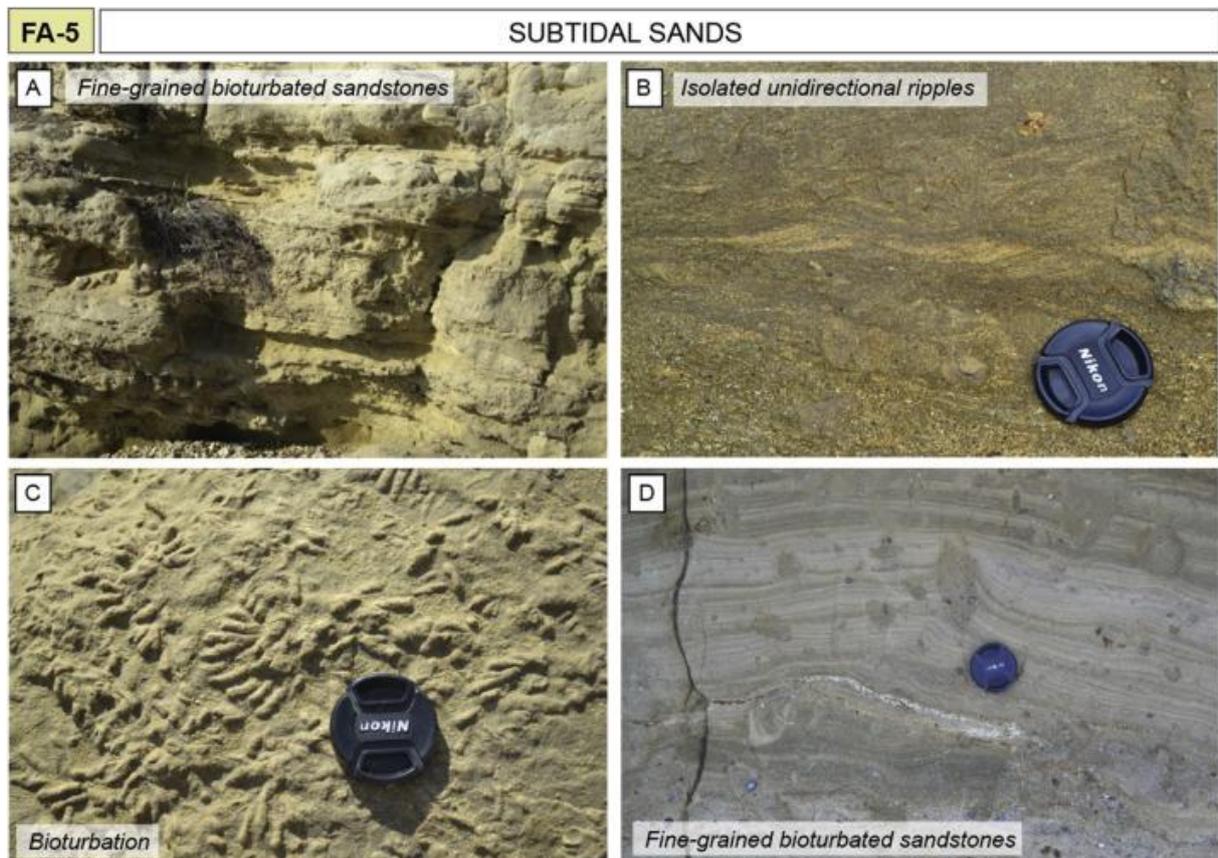


Figure 11

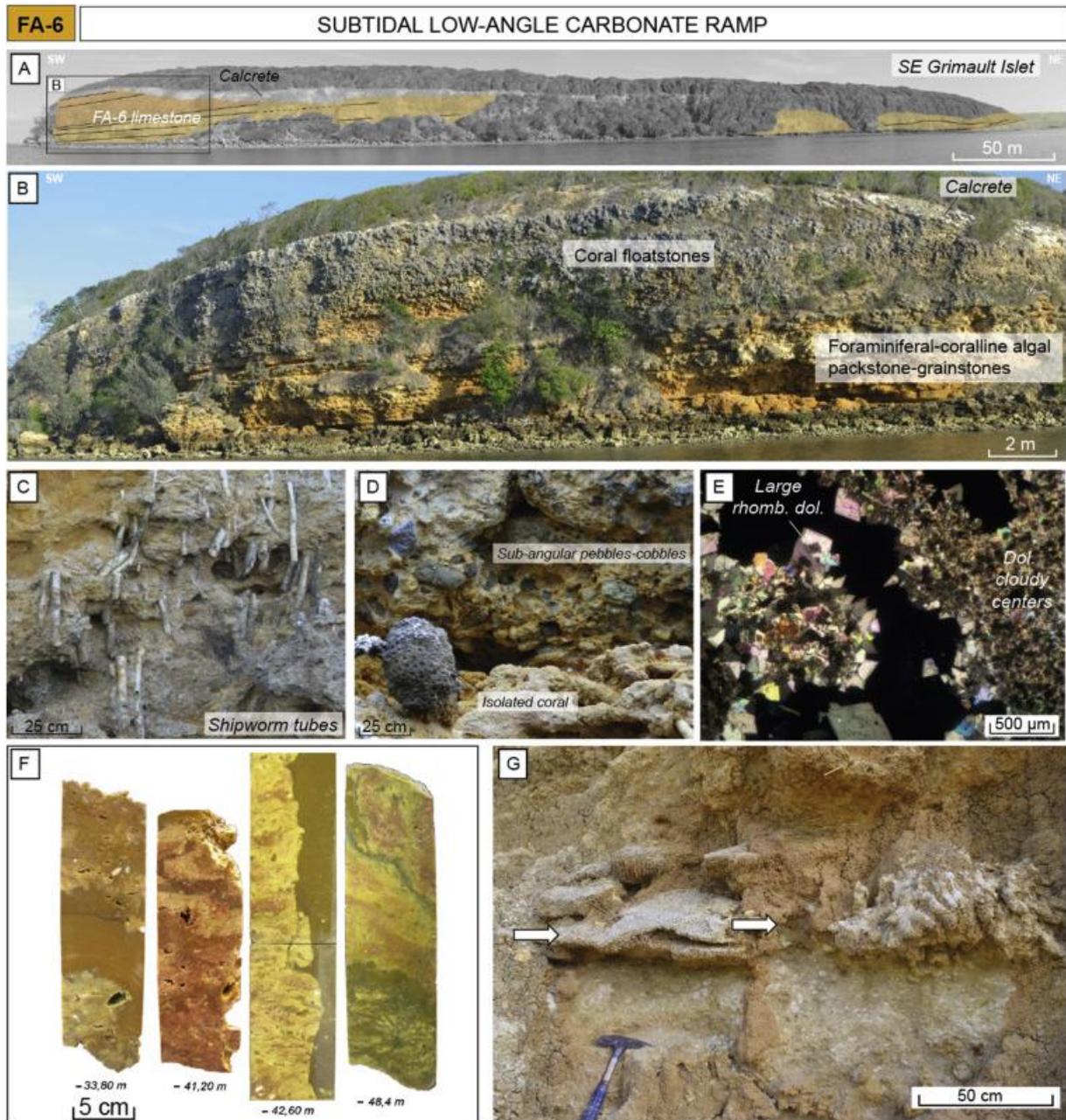


Figure 12

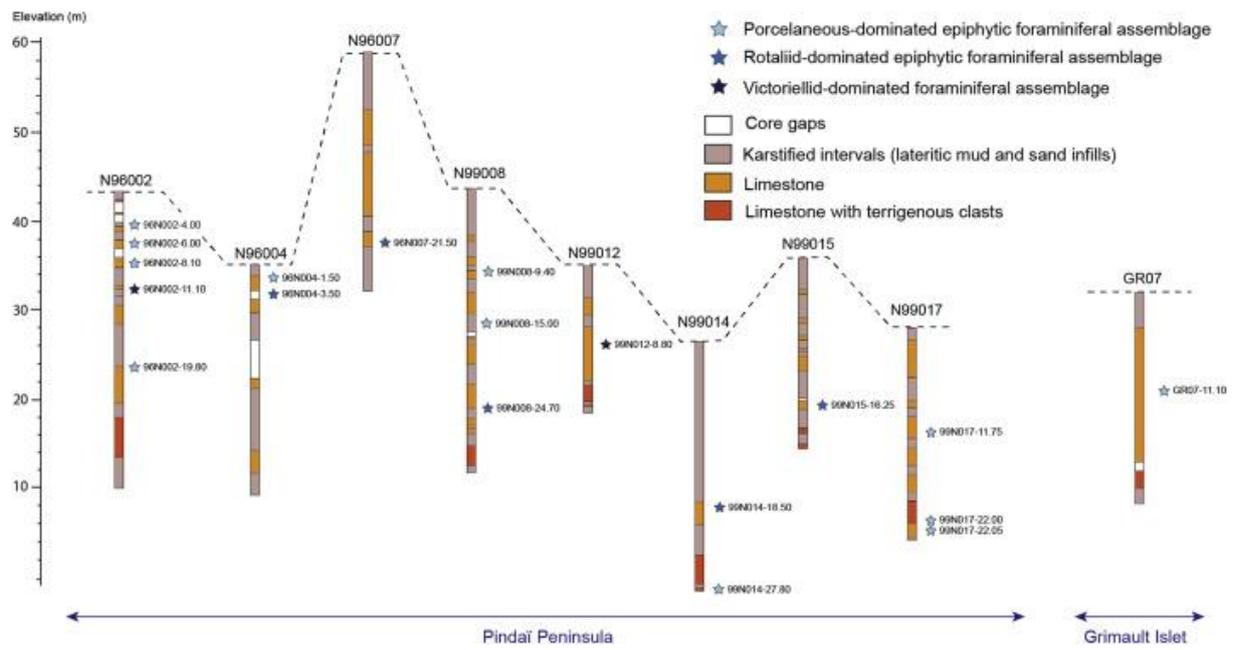


Figure 13

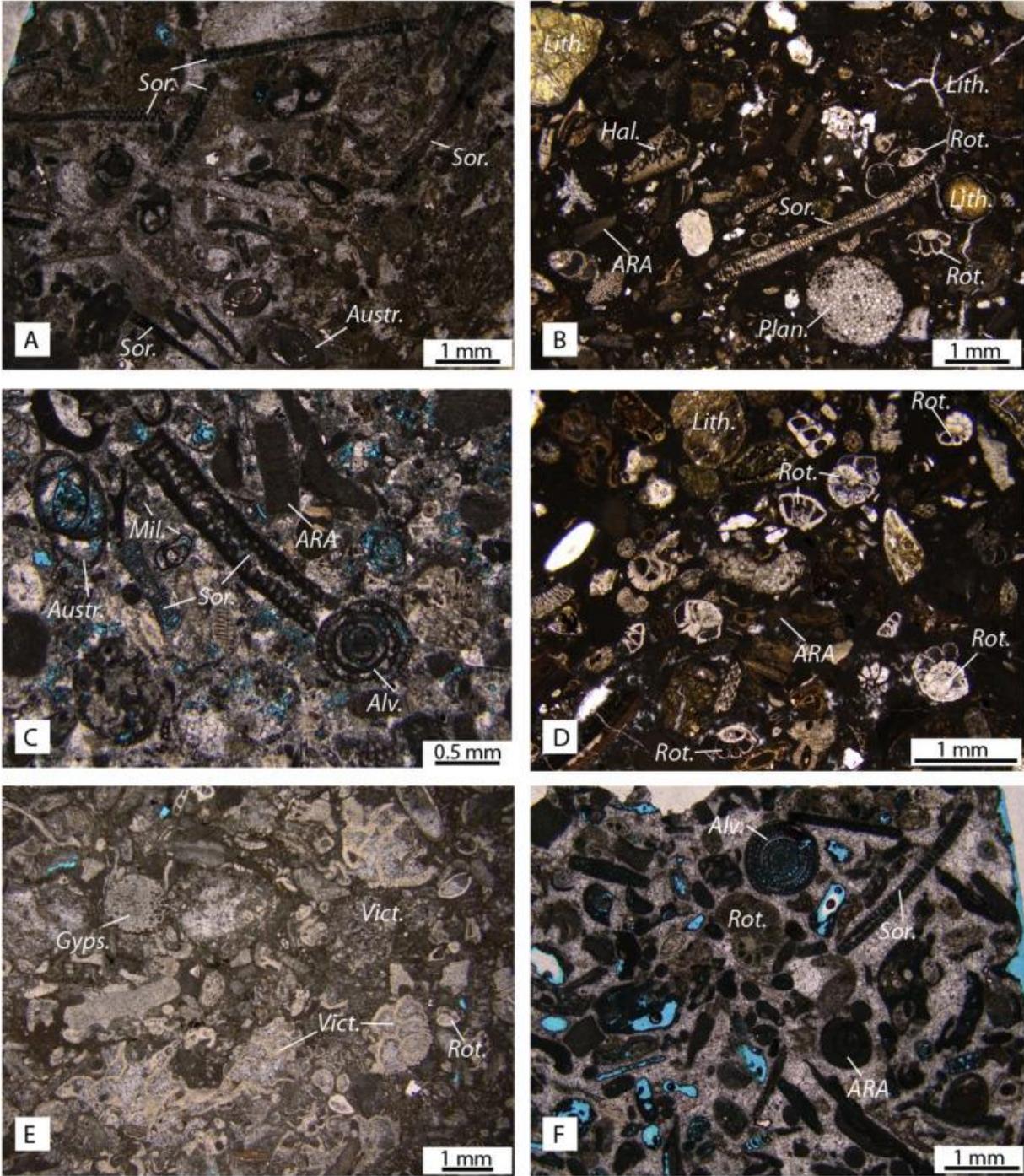


Figure 14

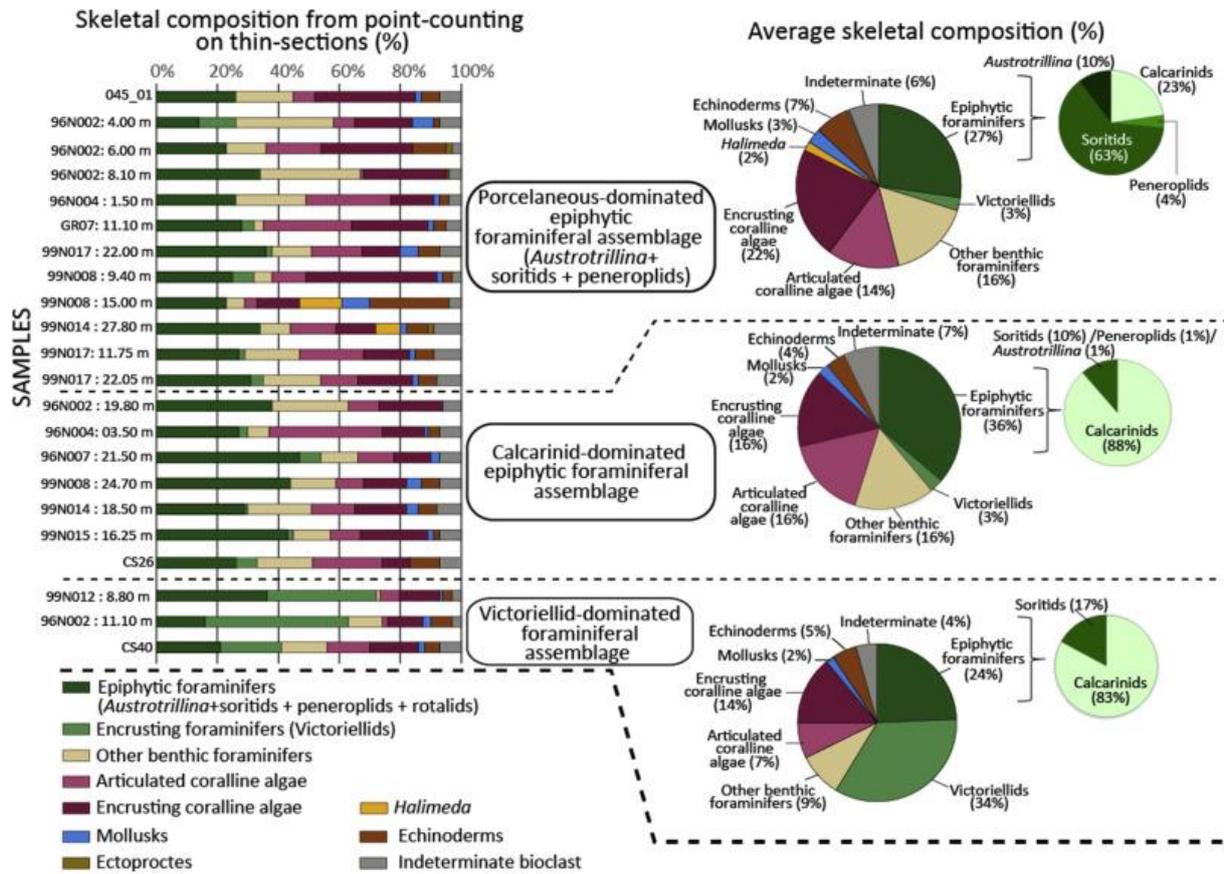


Figure 15

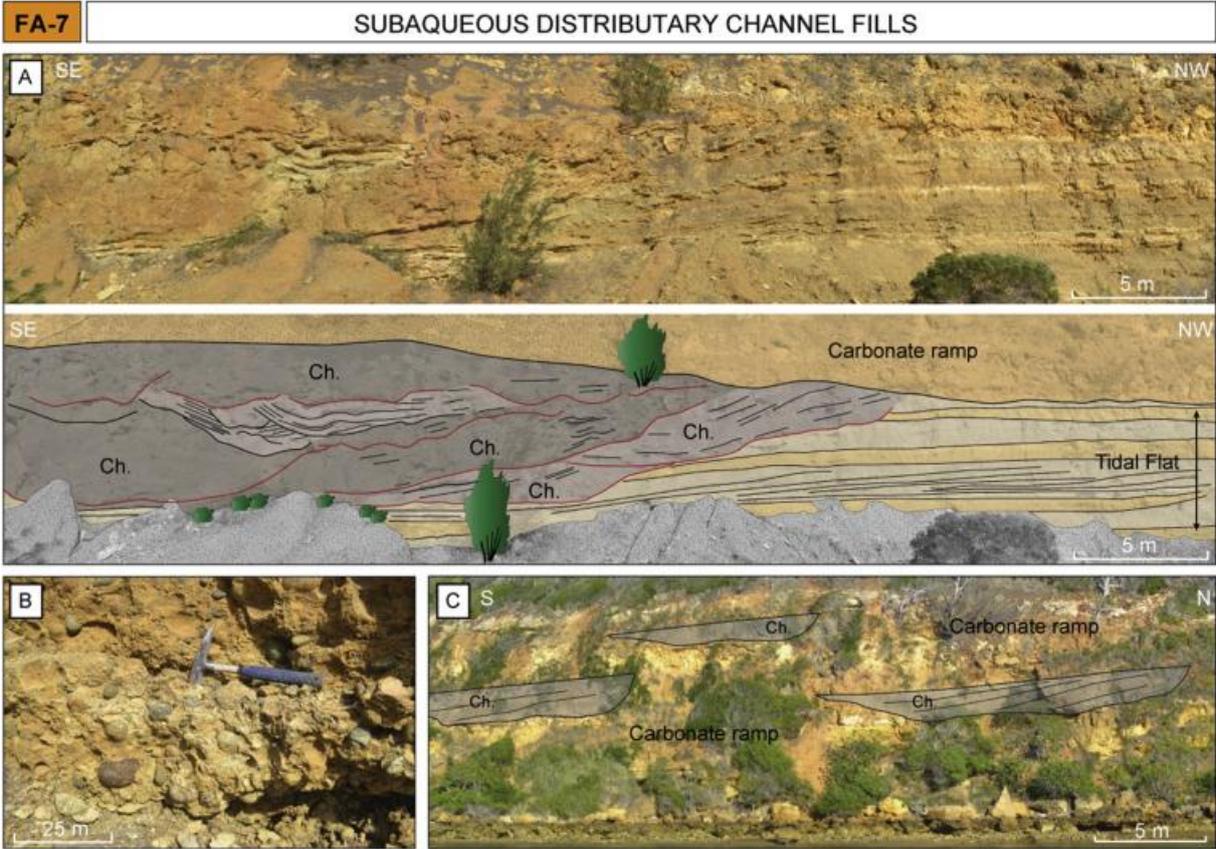


Figure 16

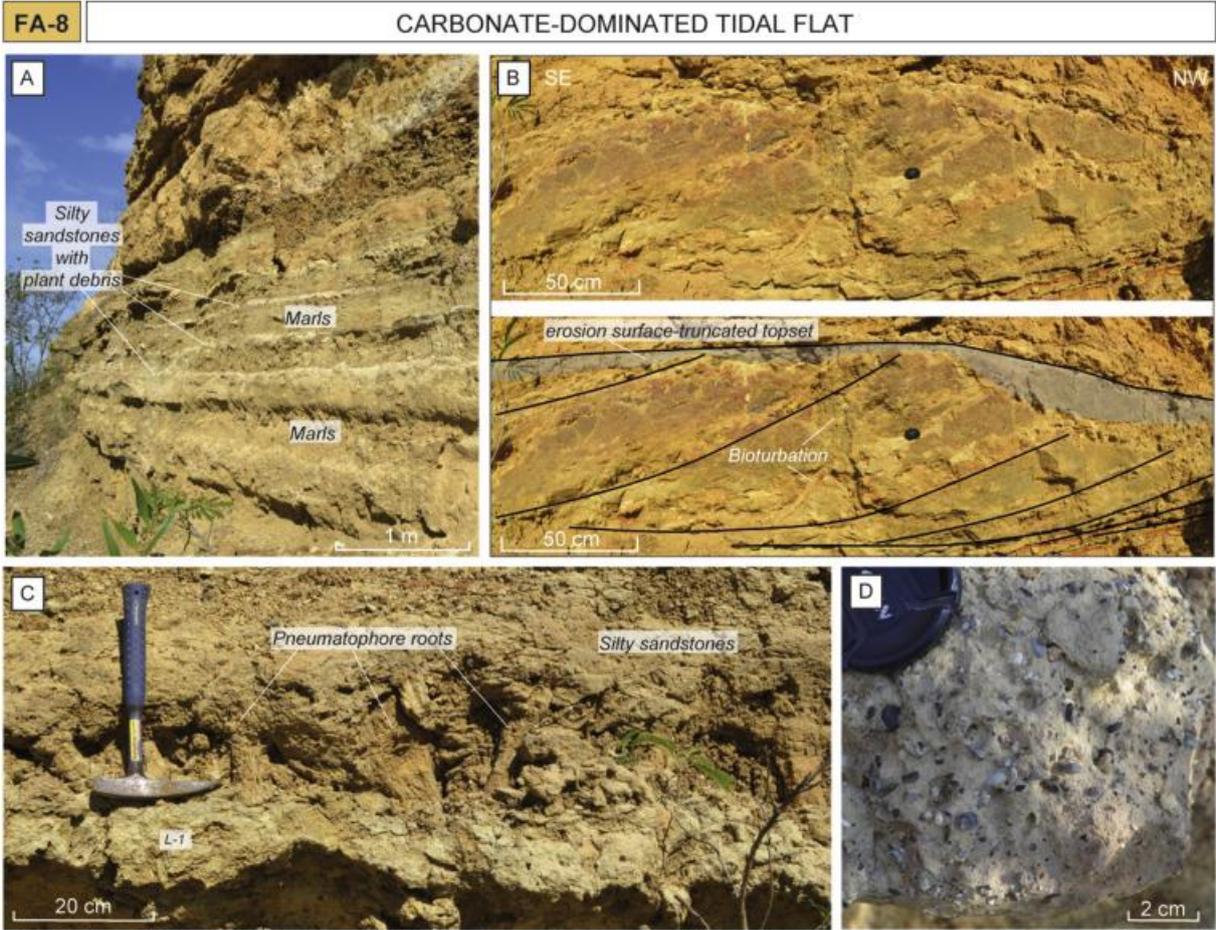


Figure 17

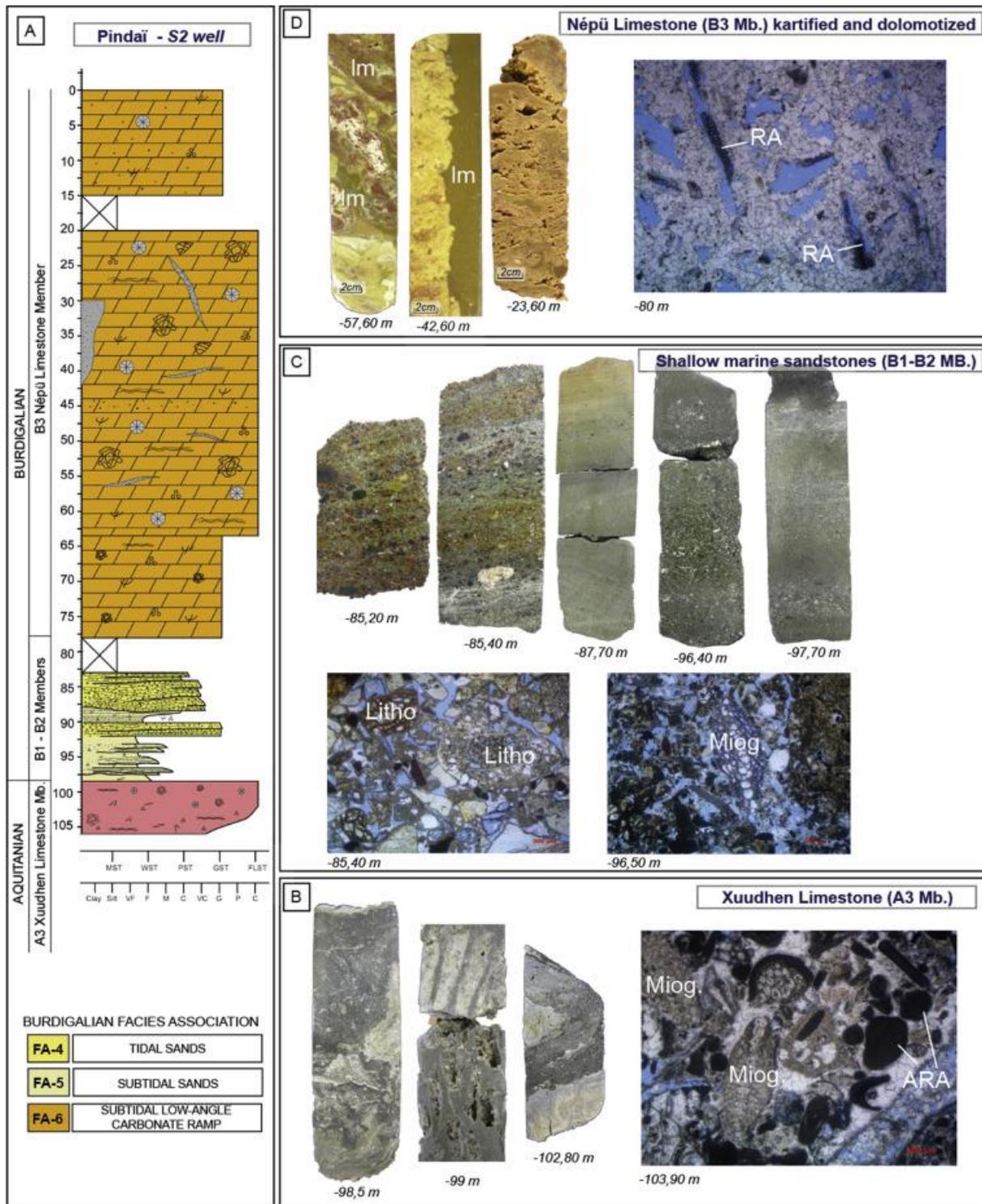


Figure 18

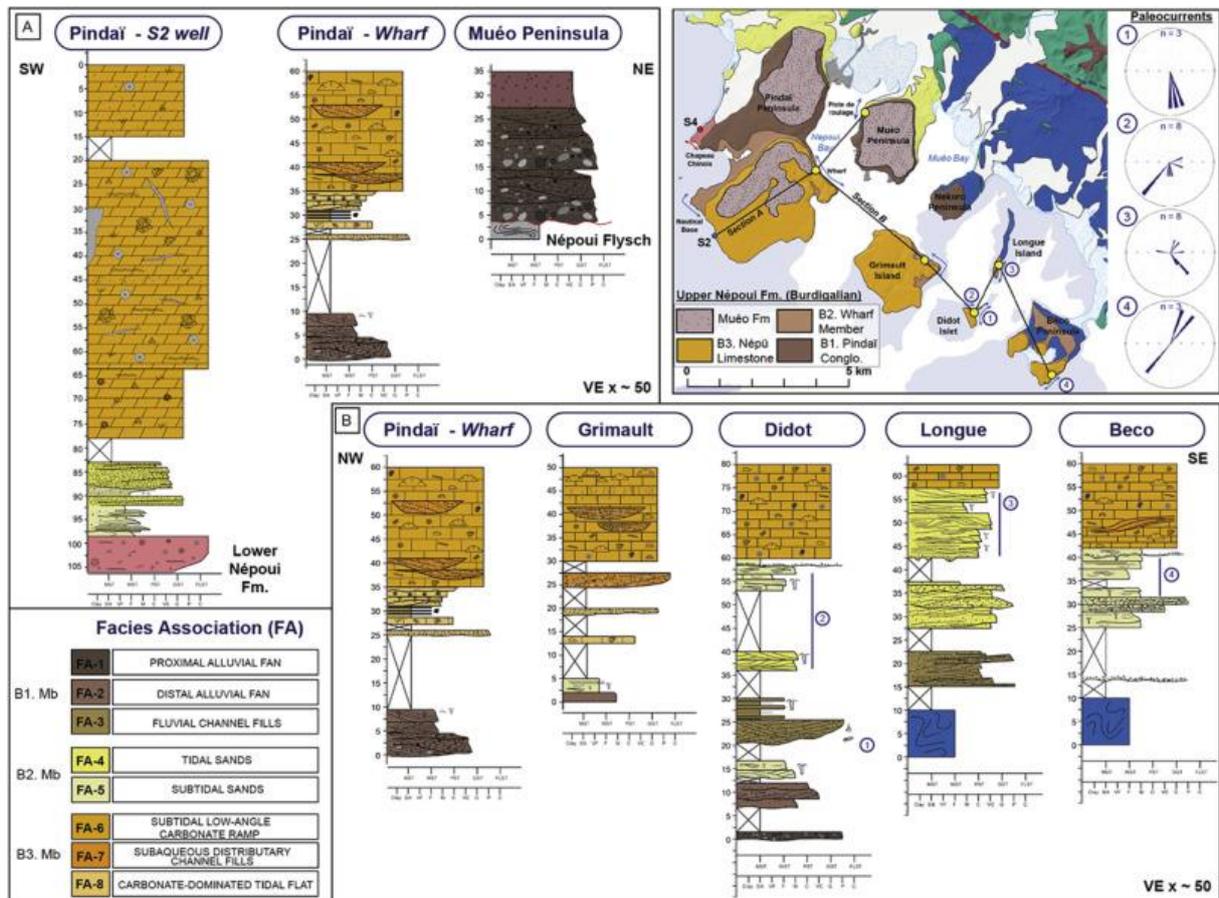


Figure 19

