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Barbara Claussmann, Julien Bailleul, Frank Chanier, Vincent Caron, Adam D Mcarthur, et al.. Contrasting mixed siliciclastic-carbonate shelf-derived gravity-driven systems in compressional intra-slope basins (southern Hikurangi margin, New Zealand). *Marine and Petroleum Geology*, 2021, 134, pp.105252. 10.1016/j.marpetgeo.2021.105252 . hal-03321847

HAL Id: hal-03321847

<https://hal.science/hal-03321847>

Submitted on 18 Aug 2021

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1 **Contrasting mixed siliciclastic-carbonate shelf-derived gravity-**
2 **driven systems in compressional intra-slope basins**
3 **(southern Hikurangi margin, New Zealand)**

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15 ***Accepted Manuscript version (unedited PDF accepted for publication).***

16 ***Published Journal Article version available online, 30 July 2021:***

17 ***<https://doi.org/10.1016/j.marpetgeo.2021.105252>***

18
19 ***Published by Elsevier in Marine and Petroleum Geology Journal.***

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22 1. ABSTRACT

23 Along active margins, the combination of predominant tectonic activity and shallow-marine mixed
24 siliciclastic-carbonate source systems developing upon and around actively growing structures
25 challenges traditional source-to-sink models. This study aims to investigate the implications of mixed
26 siliciclastic-carbonate shelfal domains located in contrasting geotectonic settings (thrust forelimb and
27 backlimb) for the development of the concomitant gravity-driven systems beyond the shelf edges. Here,
28 we document the vertical and lateral stratigraphic variabilities of the shelf-derived turbidites and mass-
29 transport deposits (MTDs) at outcrop-scale through the integrated interpretation of photogrammetry, field
30 and taphonomic data from the emerged southern portion of the Hikurangi subduction margin. Results
31 highlight the role and importance of varying structural setting of the sediment source, whereby the
32 different morphologies of the source regions (continent-attached forelimb, continent-detached backlimb)
33 control the development of highly varied shelf-derived gravity-driven depositional systems that interact
34 with the structures across the same confined intra-slope basin. The deposits are tens to a few hundred
35 of meters in thickness and have a lateral extent of several kilometers. The depositional systems are
36 characterized by durations of 1 to 2 Ma and were primarily controlled by the geometries and tectonic
37 motion of the underlying structures at the shelf edges. Shelf-derived mass-wasting systems occurred on
38 both sides of the actively growing thrust structures and were sourced from both shelfal domains that were
39 attached or detached from the continental domain. When sourced from the backlimbs however, the
40 subsequent MTDs exhibit more complex internal architectures, ultimately recording the dynamic changes
41 in slope gradient, and can therefore be used as proxies for unraveling the tectonic activity of an individual
42 structure. Our study provides new insights to better predict mixed siliciclastic-carbonate depositional
43 settings along active margins, sourced from thrust forelimb and backlimb. These results may be important
44 for deep-marine exploration and tectonostratigraphic reconstruction of fold-and-thrust belts.

45 **Keywords:** active margin, confined slope channel system, mass-wasting system, mixed siliciclastic-
46 carbonate systems, confined basins, trench-slope basins, thrust forelimb, thrust backlimb

47 2. INTRODUCTION

48 Understanding sediment transfer and distribution beyond the shelf edge and in the outboard part of
49 continental margins has been at the heart of numerous studies in the past 60 years (e.g., Bouma 1962;
50 Posamentier and Vail 1988; Richards et al. 1998; Posamentier and Kolla 2003; Sømme et al. 2009;

51 Paumard et al. 2020). Commonly, the pathways taken by sediments across the shelf margins are believed
52 to some extent control different types of sediment gravity flows and related depositional systems (Prather
53 2003; Gong et al. 2016; Paumard et al. 2020). Submarine canyons and slope channels are more likely
54 to funnel turbidity currents (Peakall and Sumner 2015), transferring continental and or shelf-derived
55 material downslope (Kuenen 1964; Mulder et al. 2017) and normally ending their course as weakly
56 confined to unconfined lobe systems (e.g., Brunt et al. 2013), whereas unconfined shelf edge failures
57 typically generate shelf-derived debris flows that are either captured on the slope or deposited on the
58 basin floor (e.g., Moscardelli and Wood 2008).

59 Along active margins, the combination of predominant tectonic activity and mixed siliciclastic-carbonate
60 source systems developing upon and around actively growing structures challenges traditional source-
61 to-sink models (Chiarella et al. 2012; Chiarella et al. 2019). Resulting deep marine sedimentation styles
62 are highly variable, intricate and not readily predictable (McArthur et al. 2019). Whilst the underlying
63 structural styles and motions fundamentally impact the morphology of continental shelves and their
64 margins, they also induce dynamic changes in slope gradient, confinement and connection to sediment
65 sources (Sømme et al. 2009). Shallow marine, mixed shelfal systems are common in active tectonic
66 settings (Puigdefàbregas et al. 1986; Chiarella et al. 2012). Typified by repeated terrigenous sediment
67 input (Chiarella et al. 2017), the shallow marine, mixed shelfal systems commonly result in intricate shelf
68 sedimentation and sediment routing systems (Chiarella et al. 2019; Moscardelli et al. 2019; Cumberpatch
69 et al. 2021). All these parameters control the dynamics of the flows from the shelf to the basin floor and
70 can be expected to have a significant impact on the development of the coeval deep-marine systems,
71 influencing (1) the types of sediment gravity flow, (2) the patterns of sediment dispersal, (3) the
72 morphologies (size, extent) of the systems and (4) the related sedimentary facies (e.g., Fisher 1983;
73 Pirmez et al. 2000; Mulder and Alexander 2001; Sprague et al. 2005; Sømme et al. 2009; Mulder et al.
74 2012; Moscardelli et al. 2019; Paumard et al. 2020).

75 Although previous studies looked at the depositional frameworks of shallow-marine mixed siliciclastic-
76 carbonate systems along tectonically active margins (e.g., Caron et al. 2004; Čosović et al. 2018;
77 Chiarella et al. 2019), the stratigraphic implications for the subsequent shelf-derived gravity-driven
78 systems remain unclear. Few studies have proposed generic depositional models linking mixed shallow
79 systems with their potential coeval deep-marine counterparts. However, they mostly rely on subsurface
80 data (e.g., Moscardelli et al. 2019) and only rarely discuss the impact of changes in structural setting of
81 the sediment source (e.g., thrust forelimb and backlimb settings) on the deposits (e.g., Čosović et al.
82 2018).

83 Here, we investigate the type and distribution of shelf-derived sediments, sourced from mixed siliciclastic-
84 carbonate systems (*i.e.*, compositional systems (*sensu* Chiarella et al. (2017))) that developed above the
85 asymmetrical basin-bounding structures (*e.g.* thrusts and or fault-growing folds) of an actively deforming,
86 confined intra-slope basin. We examine the vertical and lateral stratigraphic variability of the shelf-derived
87 gravity-driven deposits, at outcrop-scale, from the Whareama trench-slope basin, in the emerged
88 southern portion of the Hikurangi subduction margin, North Island of New Zealand. The deposits are of
89 Middle Miocene age and incorporate similar material (nature and size of the clasts), albeit transported
90 through a variety of sediment gravity flows, thereby suggesting different source, delivery and geotectonic
91 settings across the same confined intra-slope basin.

92 Our study aims to determine how variations in the structural style (thrust forelimb and backlimb) and
93 tectonic activity (renewal or quiescence) influence the generation of contrasting mixed siliciclastic-
94 carbonate shelf-derived gravity-driven systems and deposits across a confined, intra-slope basin (*e.g.*,
95 trench-slope basin).

96 Specific objectives are to:

- 97 • Document the nature, architecture, and size of the different styles of shelf-derived sedimentation
98 that can occur in a confined intra-slope basin, through an integrated interpretation of
99 photogrammetry, fieldwork and taphonomic data.
- 100 • Gain new understandings on the controls on sediment transfer and distribution beyond the edges
101 of mixed siliciclastic-carbonate shelves located upon and around the forelimb and backlimb of
102 actively growing asymmetrical structures.
- 103 • Develop generic depositional models of the distribution and styles of shelf-derived systems and
104 subsequent deposits in tectonically active settings.
- 105 • Determine the implications for the stratigraphic prediction of deep-marine, shelf-derived gravity-
106 driven systems and, of shelf-derived mass-wasting systems along active margins.

107 **3. GEOLOGICAL SETTING**

108 **3.1. Regional setting**

109 The study area is located on the eastern North Island of New Zealand, on the tectonically active Hikurangi
110 Margin (Figure 1). The Hikurangi subduction wedge (*sensu* Bailleul et al. 2013) formed in response to
111 the Uppermost Oligocene-to-Recent westward subduction of the Pacific Plate beneath the Australian

112 Plate (Ballance 1976; Spörl 1980; Pettinga 1982; Chanier and Ferrière 1991; Field et al. 1997; Nicol et
113 al. 2007). Bounded by the Hikurangi Trench to the east and the Forearc Basin to the west, the Hikurangi
114 subduction wedge comprises a succession of elongated trench-parallel sedimentary basins (*i.e.*, trench-
115 slope basins) separated by tectonically active structural ridges (Chanier and Ferrière 1991; Lewis and
116 Pettinga 1993; Bailleul et al. 2013; McArthur et al. 2019).

117 Since the onset of subduction, the Hikurangi subduction wedge underwent a polyphase tectonic history,
118 comprising three main tectonic periods (Figure 2 and references within). Complex stratigraphic
119 architectures and geometries, such as diachronous sedimentation patterns, syn-sedimentary
120 deformations and discontinuities, are characteristic of the Miocene-to-Recent trench-slope basin fills and
121 attest of the close interplay between tectonics and sedimentation (Neef 1992; Neef 1999; Bailleul et al.
122 2007; Bailleul et al. 2013; Burgreen and Graham 2014; McArthur et al. 2019; McArthur et al. 2021). Deep-
123 marine sedimentation dominates their fill and includes mass-transport deposits (MTDs), turbidites and
124 extensive hemipelagic mudstones; yet carbonate, biogenic and shallow-marine fringe deposits can also
125 be found sporadically (Figure 2) (Field et al. 1997; Lee and Begg 2002; Bailleul et al. 2007; Bland et al.
126 2015; McArthur and McCaffrey 2019). These basin fills either conformably or unconformably overlie the
127 Cretaceous to Paleogene pre-subduction basement, mainly composed of the Lower Cretaceous Torlesse
128 greywackes and the Upper Cretaceous to Oligocene detrital to pelagic series (Spörl 1980; Bradshaw
129 1989; Chanier and Ferrière 1991; Field et al. 1997; Lee and Begg 2002).

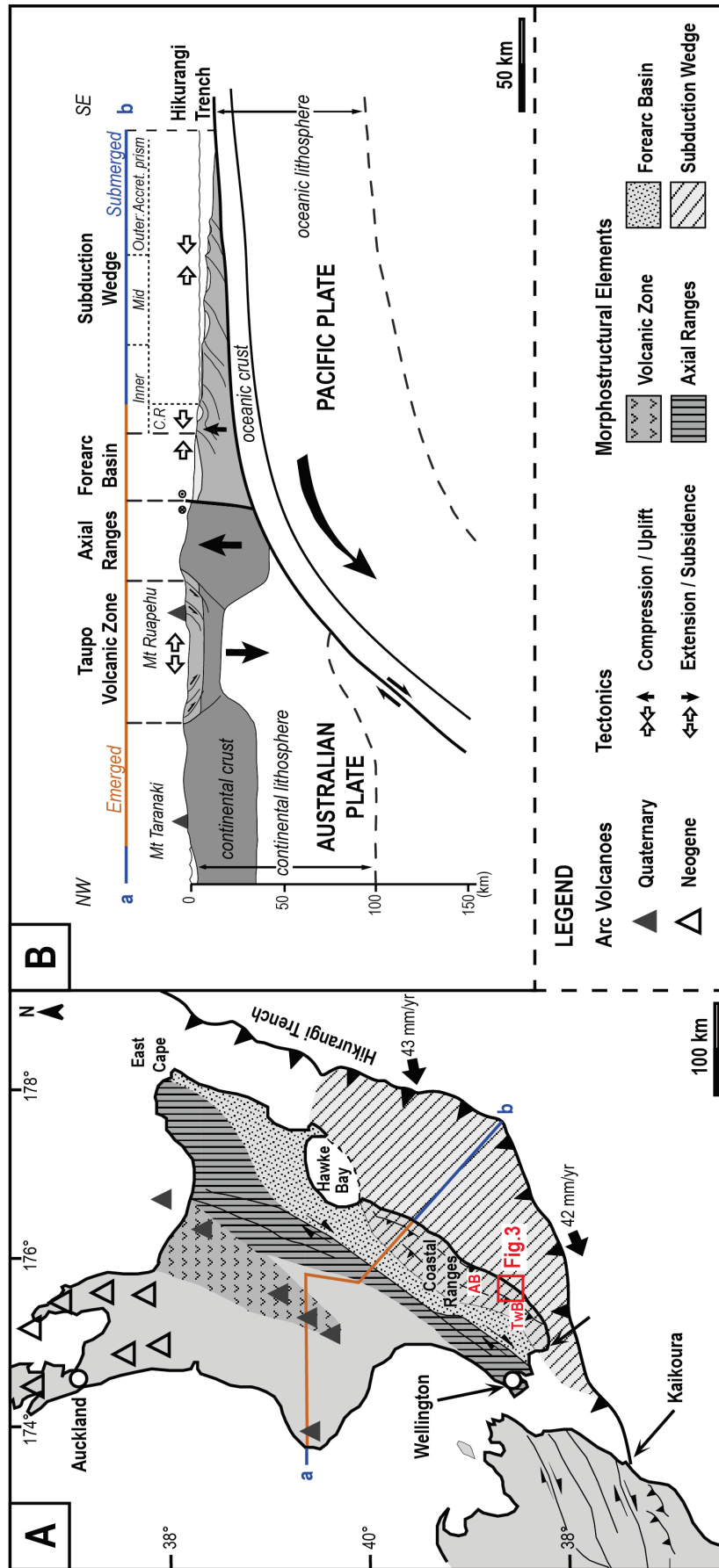


Figure 1: (A): Major subduction-related morphostructural features of the active Hikurangi Margin, north Island of New Zealand. Black arrows show present-day relative plate motion between Pacific and Australian Plates from Beavan et al. (2002). (AB – Akitio Basin; TwB – Te Wharau Basin). See (B) for the a – b general cross-section of the Hikurangi subduction complex (C.R – Coastal Ranges). Modified after Chanier et al. (1999), Baillieu et al. (2007) and Baillieu et al. (2013).

130

3.2. Southern Coastal Ranges

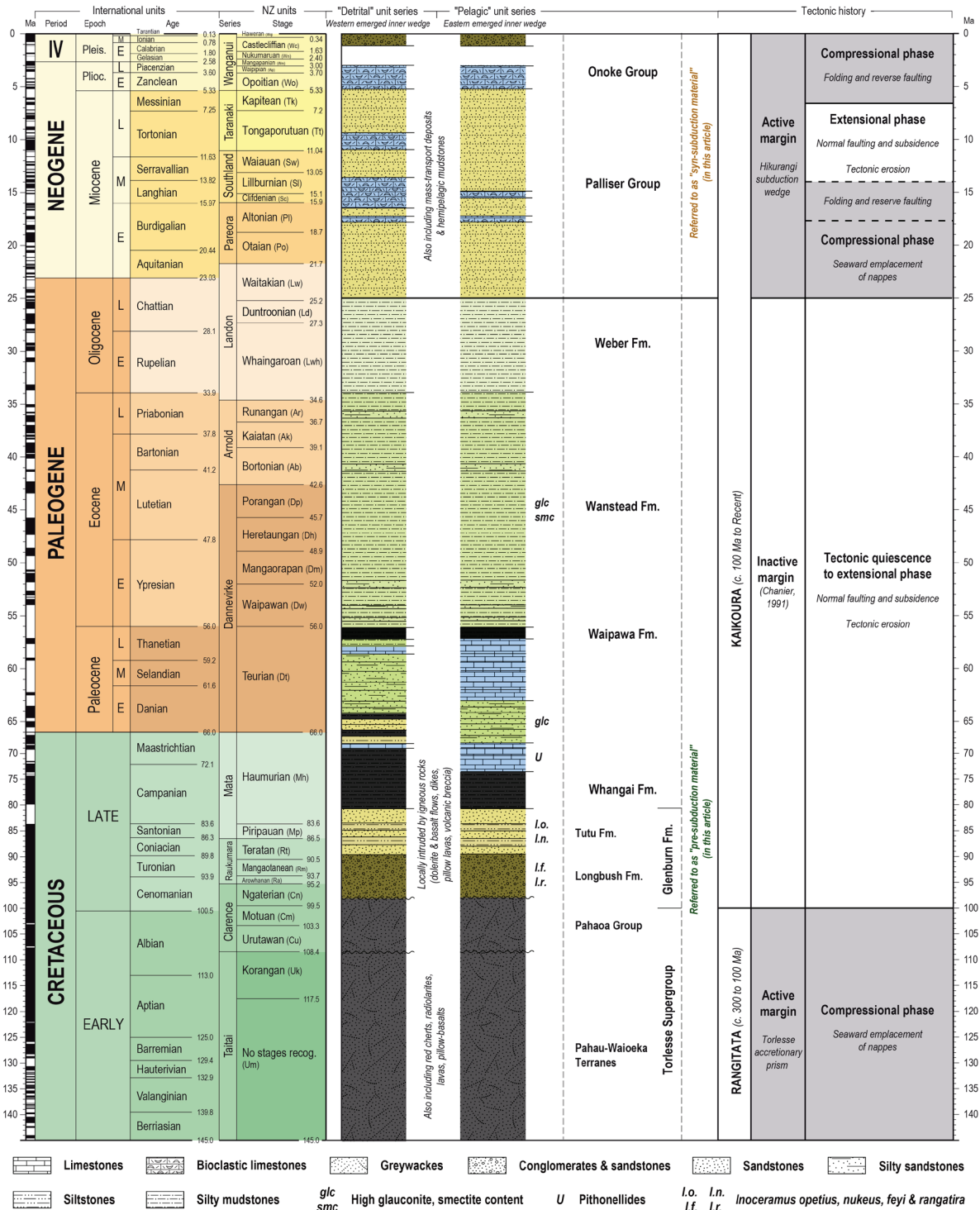
131 The study area lies within the exhumed inner portion of the subduction wedge (e.g., emergent trench-
132 slope break) that forms the southern Coastal Ranges of the eastern North Island of New Zealand. Here,
133 we focus on the tectonic settings at the end of the first compressional phase (Figure 2; Middle Miocene,
134 Clifdenian – Lillburnian (Langhian to early Serravallian, 15.9 – 13.05 Ma)), where both the inception and
135 development of the trench-slope basins occurred (Chanier and Ferrière 1991; Bailleul et al. 2013; Malie
136 et al. 2017).

137 During this time period, compressional tectonics resulted in uplift of some of the fold and thrust basin-
138 bounding structures and more generally of the margin (Crundwell 1987; Bailleul et al. 2013). The induced
139 changes in depositional environments (e.g., neritic conditions) atop the structures controlled the
140 development of shelfal environments, and, in particular, mixed siliciclastic-carbonate systems (i.e.,
141 compositional systems (*sensu* Chiarella et al. (2017)) at shallower waters whilst gravity-systems
142 dominated elsewhere at deeper waters (Crundwell 1987; Chanier and Ferrière 1991; Neef 1992; Lee and
143 Begg 2002; Bailleul et al. 2007; Bailleul et al. 2013; McArthur et al. 2019; McArthur and McCaffrey 2019;
144 Crisóstomo-Figueroa et al. 2020; McArthur et al. 2021).

145 Numerous mixed siliciclastic-carbonate systems co-existing during this period formed a regional shelf
146 domain throughout the south-western portion of Hikurangi Margin (Crundwell 1987; Chanier 1991;
147 Bailleul et al. 2007; Bailleul et al. 2013; Bailleul et al. 2015; Clausmann et al. 2021). The shelves
148 developed above substantially different substrata inherited from local tectonics, and they either
149 unconformably overlie Lower to Middle Miocene syn-subduction sedimentary rocks or Cretaceous to
150 Paleogene pre-subduction basement. Their depositional settings preferentially suggest narrow,
151 continent-attached shelfal systems receiving regular terrigenous inputs from the hinterland (Bailleul et al.
152 2015; Clausmann et al. 2021). However, isolated, continent-detached platform systems may also have
153 locally occurred on some of the actively growing structures (see (Caron et al. 2004; Caron et al. 2021)
154 for examples along the same margin, yet at different geological times).

155 Clausmann et al. (2021) previously demonstrated that short-lived periods (ca. 1 to 2 Ma) of sustained
156 (with or without phases of quiescence) compressional tectonics favor the propagation of the folds and
157 thrusts, and thus the expansion of abrupt, unstable areas close to these structurally-controlled shelf-
158 margins. Repeated uplift and slope oversteepening of these areas eventually lead to recurrent shelf
159 collapses, sourcing shelf-derived sediments (e.g., mass-wasting products) into deep water. The
160 associated deposits result from sediment gravity flows, such as cohesive debris flows or mudflows, and

161 rework a substantial amount of shelf-derived material, incorporating sediments and fossils from neritic
 162 shelfal environments. Several occurrences of such shelf-derived mass-wasting events and related
 163 deposits of Middle Miocene age were identified in the southern Coastal Ranges cropping out across
 164 several trench-slope basins (*i.e.*, Whareama, Te Wharau and Akitio Basins) (Figure 1; Figure 2; Figure
 165 3).



166

167 Figure 2: [previous page] Chronostratigraphic chart for the southern, emerged portion of the Hikurangi subduction wedge
168 adapted from [Claussmann et al. \(2021\)](#). Lithostratigraphy details adapted from [Chanier et al. \(1990\)](#), [Chanier \(1991\)](#), [Chanier](#)
169 [and Ferrière \(1991\)](#), [Field et al. \(1997\)](#), [Lee and Begg \(2002\)](#) and [Bland et al. \(2015\)](#) and detailing the pre- and syn-Hikurangi
170 subduction series. Regional tectonism adapted from [Chanier et al. \(1999\)](#), [Bailleul et al. \(2013\)](#) and [Malie et al. \(2017\)](#). New
171 Zealand stages after [Raine et al. \(2015\)](#) showing the equivalence with the international stages.

172 **3.3. The Whareama Basin**

173 The Whareama Basin is narrow (two to six kilometers wide), elongated (50 kilometers long) and trench-
174 parallel (NE-SW) ([Johnston 1980](#); [Chanier 1991](#)). It is bounded by pre-Miocene basement ridges that are
175 controlled by the Adams-Tinui Fault complex and Pukeroro Fault to the west (*i.e.*, landward basin margins,
176 Figure 3) and the Flat Point-Whakataki Fault complex to the east (*i.e.*, seaward basin margin, Figure 3).
177 The basin's development began at the onset of subduction and thus records gravity-driven deposits
178 dating from the earliest Miocene. It formed on the back of the Glenburn Nappe, a trenchward-
179 advancing thrust sheet composed of Cretaceous to Eocene pre-subduction series that are either
180 unconformably or conformably overlain by syn-subduction deposits ([Chanier 1991](#); [Chanier and Ferrière](#)
181 [1991](#)).

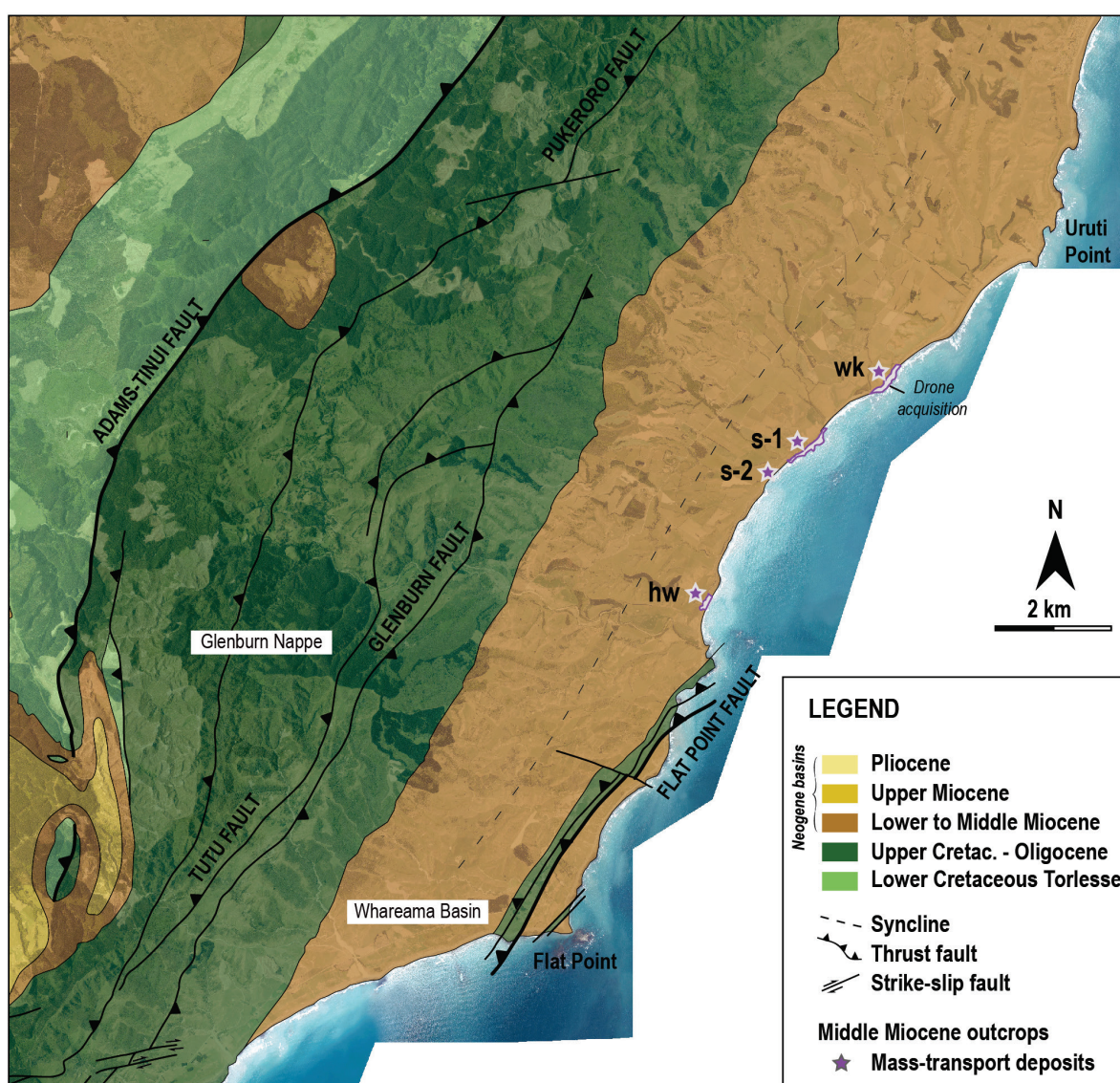
182 The study area is located on the eastern limb of the Whareama Basin, with beds strongly dipping towards
183 the west (ca. 40°). Nearby, well-exposed Middle Miocene shelf-derived MTDs were previously described
184 and attributed to result from the collapses of coevally developing narrow, continent-attached shelf-
185 margins located to the north-west (Figure 3; Sefton Hills s-1 and s-2 sections) ([Claussmann et al. 2021](#)).

186 **4. DATA AND METHODS**

187 This outcrop-based study integrates traditional and digital field data acquired during three field campaigns
188 (2018, 2019, 2020) in the southern portion of the Hikurangi subduction wedge (Coastal Ranges, eastern
189 North Island of New Zealand), and focuses on the shelf-derived gravity-driven deposits of the Waikaraka
190 and Homewood localities (Figure 3; wk and hw sections).

191 Fieldwork consisted of logging sedimentological profiles (total of 312 meters at 1:50 bed-scale),
192 measuring structural (*e.g.*, ductile deformation analysis) and stratigraphic features (*e.g.*, bedding,
193 paleocurrent) as well as describing each shelf-derived flow deposit in a standardized manner (facies and
194 geometries).

195 Owing to the abundant macrofossils contained in the deposits, occurring both as whole skeletons and
 196 fragmented remains, five taphonomic analyses were performed on approximately one square meter
 197 areas for both the outcrops (*i.e.*, four at Waikaraka and one at Homewood). These analyses were
 198 restricted to the coarsest fraction (larger than five millimeters) of death assemblages contained in the
 199 deposits. Three categories of skeleton damages, namely fragmentation, abrasion and bioerosion, were
 200 described visually using the graded classification scale presented in Appendix 1, complemented by Caron
 201 (2011) and Caron et al. (2019).



202
 203 Figure 3: Satellite map from World Imagery (ESRI), and onshore geological map from Chanier (1991) of the Whareama Basin.
 204 Location of the drone acquisition and related 3D outcrop models presented in this study = wk = Waikaraka section and hw =
 205 Homewood section. See Claussmann et al. (2021) for the s-1 and s-2: Sefton Hills section and 3D outcrop model.

206 Seven samples were collected to supplement the Fossil Record Electronic Database (<https://fred.org.nz/>)
207 since both areas are currently mapped as Quaternary beach deposits on the 1:250 000 Geological Map
208 of New Zealand (QMAP) of Wairarapa (<https://www.gns.cri.nz/>). The micro- and macropaleontological
209 analyses were conducted by GNS New Zealand and allowed to determine the age of the sedimentary
210 units and related depositional paleobathymetries (Appendix 2; Appendix 3).

211 Digital fieldwork involved the acquisition of high-resolution aerial images (1,900 at Waikaraka and 690 at
212 Homewood) using a DJI Phantom 4 Pro drone, supplemented by a number of Ground Control Points
213 (GCP) collected with a Trimble GeoExplorer 2008 differential global positioning system tool (DGPS) and
214 a Trimble Tempest antenna positioned at two meters above the measured point. Two georeferenced 3D
215 outcrop models were then created in the form of high-resolution triangulated mesh textured with the
216 photographs.

217 5. FACIES AND ARCHITECTURAL ELEMENTS

218 5.1. Facies and architectural schemes

219 Integration of the data allowed an analysis and comparison of the stratigraphic architectures and facies
220 organization within both outcrops.

221 A total of 14 lithofacies were recognized across the two outcrops (summarized in Table 1 and illustrated
222 in Figure 4, Figure 5, Figure 6 and Figure 7) and classified according to their dominant lithology, primary
223 sedimentary features and interpreted in terms of flow processes using models from (Bouma 1962; Nardin
224 et al. 1979; Lowe 1982; Postma 1986; Kneller 1995; Mulder and Alexander 2001).

225 These comprise massive silty mudstones (**MDST**); tabular, very thin- to medium-bedded (**LDTC-a**) or
226 medium- to thick-bedded (**LDTC-b**) sandstones with mudstone caps; lenticular, medium to very-thick
227 bedded sandstones (**HDTC**); structured (**SST-a**) or structureless (**SST-b**) sandstones; pebbly sandstones
228 (**sHDTC-a**); bioclastic grits (**sHDTC-b**); lenticular, medium to very-thick bedded gravel sandstones with
229 mudstone caps (**gsHDTC**); organized to disorganized clast-supported (respectively **gHDTC**, **CF**) and
230 matrix-support (**DF**) conglomerates; as well as deformed (**SL-a**) or undeformed (**SL-b**) mass of
231 sediments.

Deposit	Dominant lithology	Lithofacies code	Thickness (cm)	Lithofacies description	Classification and process interpretation
Turbidite	Mudstone	MDST Massive, siltstone to silty mudstone	1 to 60 (HMD) nul (WKS)	Siltstone to silty mudstone. Mainly massive, rare parallel lamination visible. Ungraded to (normally) graded. Variable degree of bioturbation, tends to be highly bioturbated.	(Td)-Te Low-density turbidity currents and hemipelagic suspension settling
	Sandstone to Mudstone	LDTC-a Tabular, very thin-to medium-bedded sandstone with mudstone cap	1 to 20 (HMD) 2 to 20 (WKS)	Fine- to medium-grained. Parallel laminations (Tb), passing into climbing ripples (Tc) with sometimes dewatering or soft sedimentation deformation structures installed above. Sometimes only developing the parallel laminations or sometimes directly developing the climbing ripples. Laminations can be highlighted by organic-rich, carbonaceous debris and shell ashes. Common bioturbation.	Tb-e Tb-Te Tc-Te Low-density turbidity currents
		LDTC-b Tabular, medium to thick-bedded sandstone with mudstone cap	5 to 100 (HMD) 10 to 100 (WKS)	Fine- to coarse-grained. If present, medium- to very coarse-grained Ta at the base of the thicker beds, commonly massive and structureless, with sand- to granule-(up to pebble-) grade clasts (lithoclasts, bioclasts and rip-up mudstone clasts), sometimes displaying sole marks such as flute casts. Fine- to medium-grained, well-developed parallel laminations (Tb), climbing ripples (Tc) with often dewatering or soft sedimentation deformation structures installed above. Common amalgamation. Lamination often highlighted by organic-rich, carbonaceous debris and shell ashes. Common bioturbation.	Ta-e Low-density turbidity currents
		HDTC Lenticular, medium to very thick-bedded sandstone with mudstone cap	5 to 20 (HMD) 5 to 200 (WKS)	Fine- to coarse-grained. Lenticular, medium- to very-coarse-grained base, commonly massive and structureless, with sand- to granule-(up to pebble-) grade clasts (lithoclasts, bioclasts and (elongated) rip-up mudstone clasts), overlain by well-developed parallel or cross-laminations in the same material, overall normally graded. Followed by fine- to medium-grained, well-developed parallel laminations (Tb), climbing ripples (Tc) with often dewatering or soft sedimentation deformation structures installed above. Lamination often highlighted by organic-rich, carbonaceous debris and shell ashes. Common bioturbation. HMD only presents the bottom part.	S-Ta-e Sandy high- to low-density turbidity currents
	Sandstone	SST-a Structured sandstone, above gHDTC, CF, DF or MF	5 to 20 (HMD) 15 to 100 (WKS)	Fine- to medium-grained. Parallel, sometimes passing into faint climbing ripples and or dewatering or soft sedimentation deformation structures, usually normally graded. Common scattered shell ashes; organic-rich, carbonaceous debris, and or granule- (to pebble-) grade lithoclasts. Mudstone cap commonly absent.	Tb-Tc Turbulent cloud, deposition from traction/traction-plus-fallout
		SST-b Structureless sandstone, above gHDTC, CF, DF or MF	5 to 30 (HMD) 15 to 100 (WKS)	Fine- to coarse-grained. Structureless, sometimes dewatering structures. Ungraded or (normally) graded. Common scattered shell ashes; organic-rich, carbonaceous debris, and or granule- (to pebble-) grade lithoclasts. Mudstone cap commonly absent.	S Turbulent cloud, rapid suspension fall-out
	Conglomerate to Sandstone	sHDTC-a Pebbly sandstone	5 to 30 (HMD) nul (WKS)	Fine- to coarse-grained. Structureless, or crude parallel laminations, commonly normally graded. Presenting extraformational clasts and rip-up mudstone clasts scattered or organized. Common mudstone cap.	S-Te Sandy high-density turbidity currents
		sHDTC-b Dis- to organized clast-supported, bioclastic	1 to 20 (HMD) nul (WKS)	Clast-supported, siltstones to silty mudstones with ≤90% of granule- to pebble-grade, extra- and intraformational clasts. Ungraded, sometimes normally graded laterally. Highly bioclastic with possible shell ashes organization. Frequent rip-up mudstone clasts in the normally graded sections. Often overlying sand with parallel laminations and common load structures, and either overlain by structureless sand and or mudstone cap, rarely bioturbated.	R?-S/Ta?-Te Cohesionless debris flow or bed-load transport underneath gravelly to sandy high-density turbidity currents
		gsHDTC Organized gravel to fine-grained sandstone with mudstone cap	nul (HMD) 50 to 300 (WKS)	Combination of facies gHDTC (sometimes even facies CF) at the base, overlain by facies HDTC. Clast-supported interval particularly rich in shell ashes and debris content, mostly composing the binding link between the lithoclasts; frequent rip-up mudstone clasts; very frequent boulder-graded clasts marking R2-R3 transition; low-angle, lenticular, slightly erosive base. Common amalgamation in HDTC interval.	(R2)-R3-S/Ta-e Gravelly to sandy (bipartite) high-density turbidity currents: bed-load transport underneath high-density turbidity currents
	Conglomerate	gHDTC Organized clast supported	10 to 100 (HMD) 10 to 200 (WKS)	Clast-supported, siltstone to silty mudstone with ≤90% of granule- to boulder-grade, extra- and intraformational clasts. Highly bioclastic when of average granule-grade clast size, with possible shell ashes organization. Graded (inverse to normally graded, or only normally graded). Occasional clast imbrication / organization and load structures. Rare to locally frequent rip-up mudstone clasts.	R2-R3 R3 Gravelly (bipartite) high-density turbidity currents: bed-load transport underneath high-density turbidity currents
Debrite		CF Disorganized clast-supported	10 to 200 (HMD) 20 to 400 (WKS)	Clast-supported, siltstone to silty mudstone with ≤90% of granule- to boulder-grade extra- and intraformational clasts. Highly bioclastic when of average granule-grade clast size. Ungraded, disorganized. Common load structures.	R1 Cohesionless debris flow or bed-load transport underneath gravelly high-density turbidity currents
	DF Disorganized matrix-supported	150 to 400 (HMD) 100 to 400 (WKS)	Matrix-supported, siltstone to silty mudstone with varying quantity of granule- to boulder-grade extra- and intraformational clasts. Ungraded, disorganized. Common recumbent folds, shear and load structures. Traction carpet possible.	/ Cohesive debris flow to cohesionless, sheared debris flow when comprising facies CF at base, top or edges	
Slide-Slump	Alternating conglomerate, sandstone and or mudstone	SL-a Deformed mass of sediments	100 to 2000 (HMD) nul (WKS)	Contorted, remobilized facies characterized by recumbent folds. Commonly sandy to silty mudstone background sediments.	/ Slump: coherent mass of sediment that moves along a glide plane, resulting in significant internal deformation
		SL-b Undeformed mass of sediments	100 to 2000 (HMD) 100 to 200 (WKS)	Coherent, remobilized facies displaying sharp truncations, slightly concave-up geometries with downlap.	/ Slide scar / scour: initiation of movement of a coherent mass of sediment along a glide plane, no internal deformation

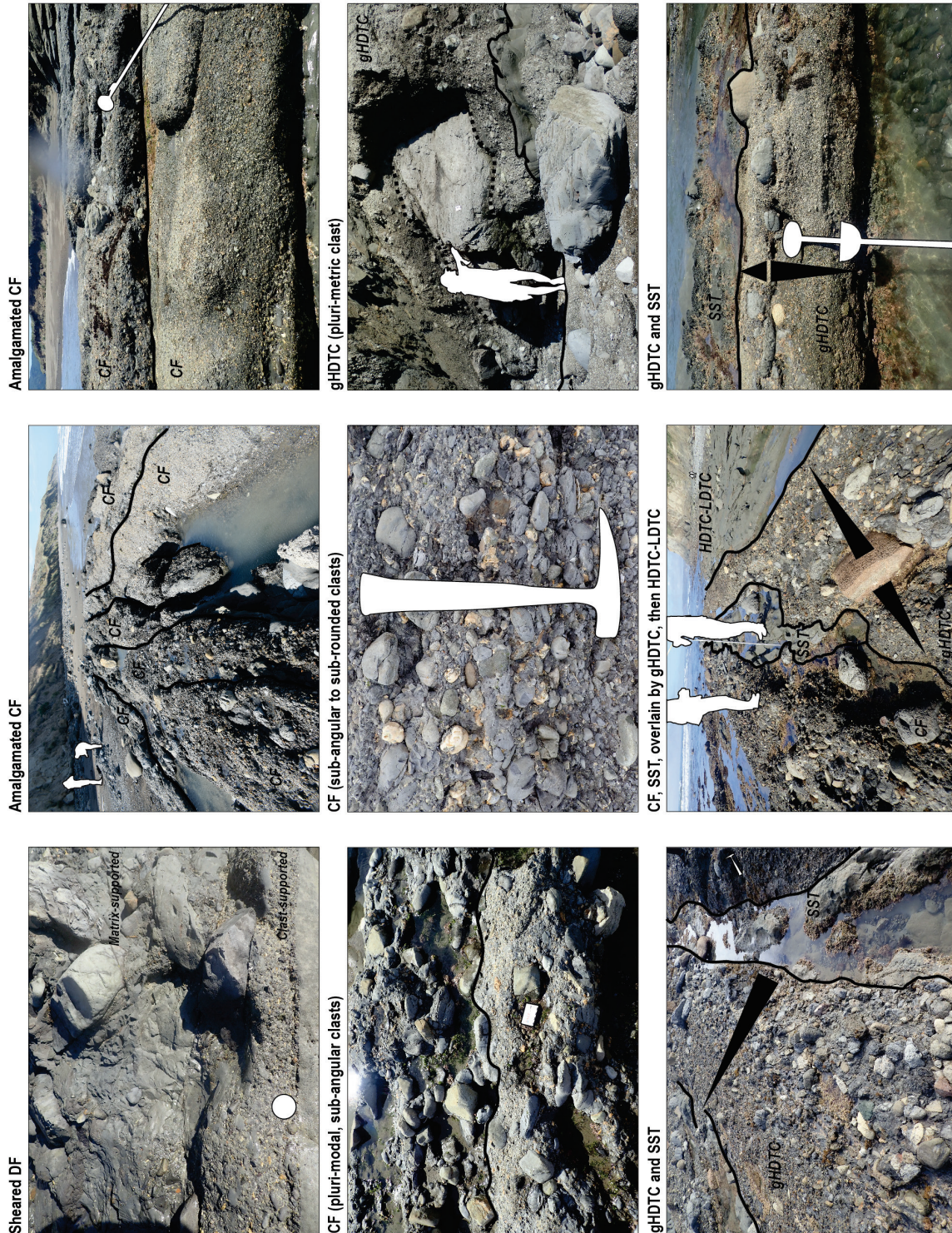
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Table 1: Lithofacies encountered at the Waikaraka (WKS) and Homewood (HMD) outcrops. See Figure 4 and Figure 5 for representative photographs of each lithofacies at Waikaraka; and Figure 6 and Figure 7 at Homewood. Note that the 'Classification' column refers to (Bouma 1962) and (Lowe 1982), and that extraformational clasts include pre- and syn-subduction lithoclasts, and syn-subduction bioclasts. Intraformational clasts include syn-subduction lithoclasts.

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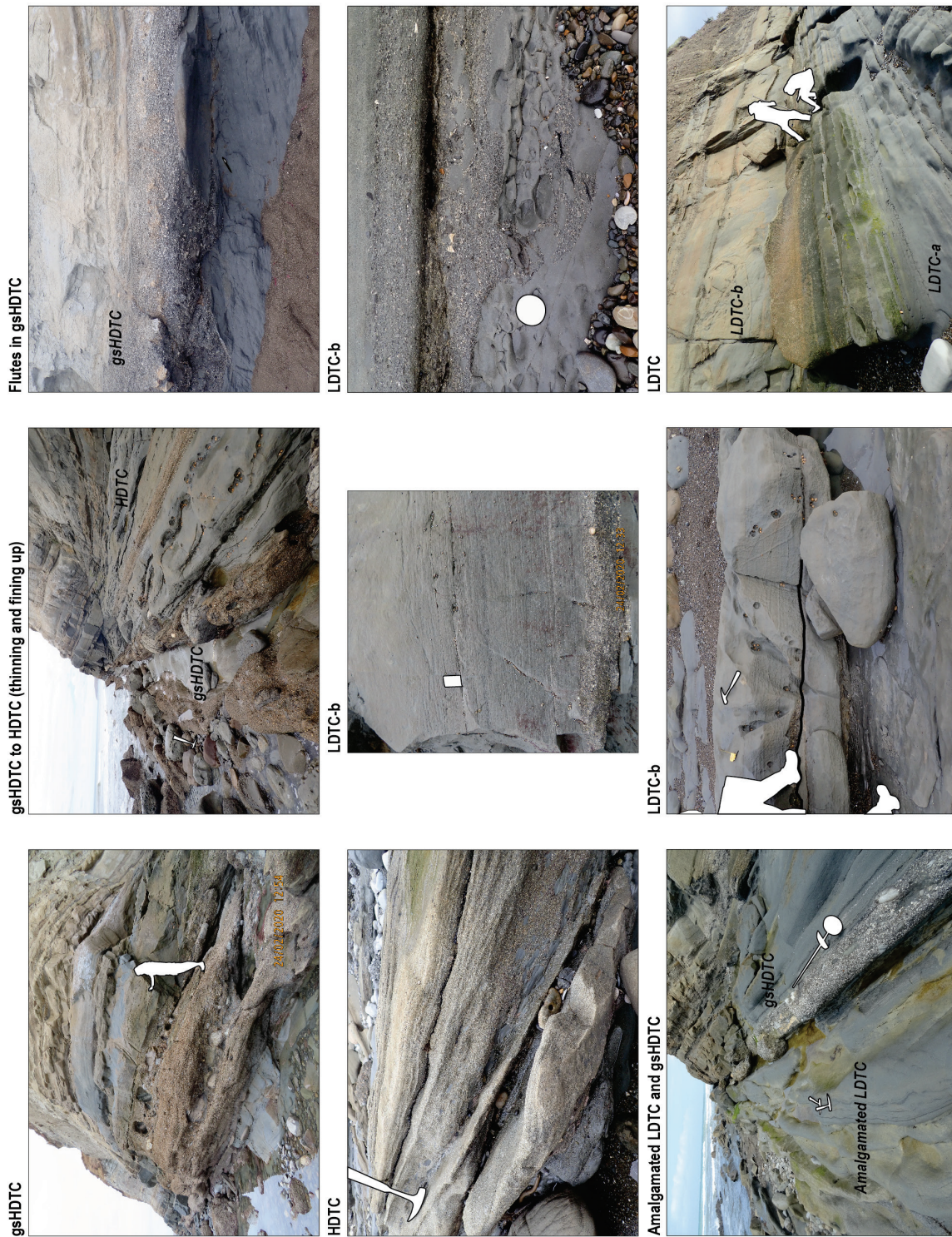
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Figure 4: Representative photographs of the lithofacies summarized in Table 1 and recognized at Waikaraka. Scales: human size (~1.70-1.75 meters); Jacob's staff (1.5 meter); hammer (33 centimeters); card (8.5 centimeters long); coin (2.3 centimeters).

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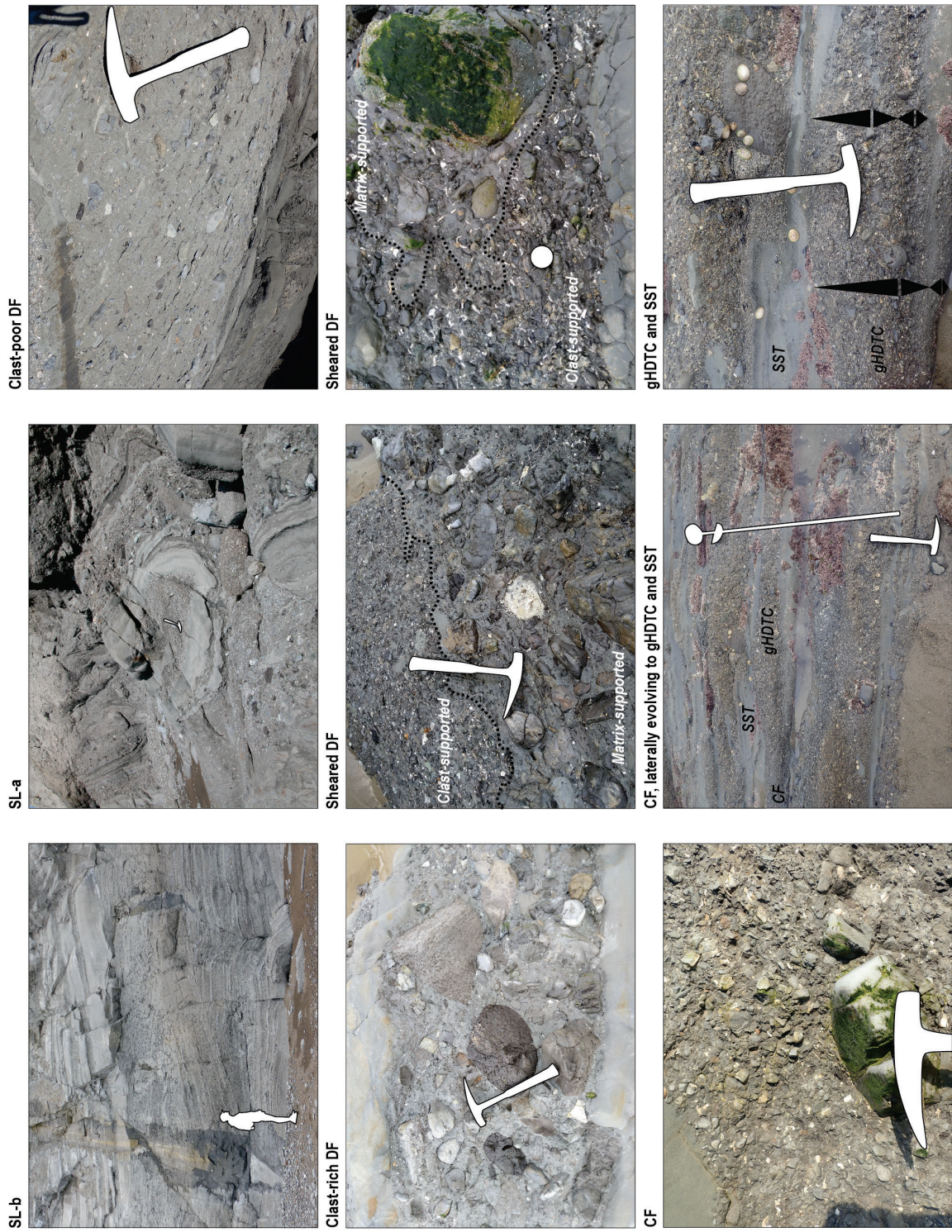


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Figure 5: Representative photographs of the lithofacies summarized in Table 1 and recognized at Waikaraka. Scales: human size (~1.70 meters); Jacob's staff (1.5 meter); hammer (33 centimeters); card (8.5 centimeters long); coin (2.5 centimeters).

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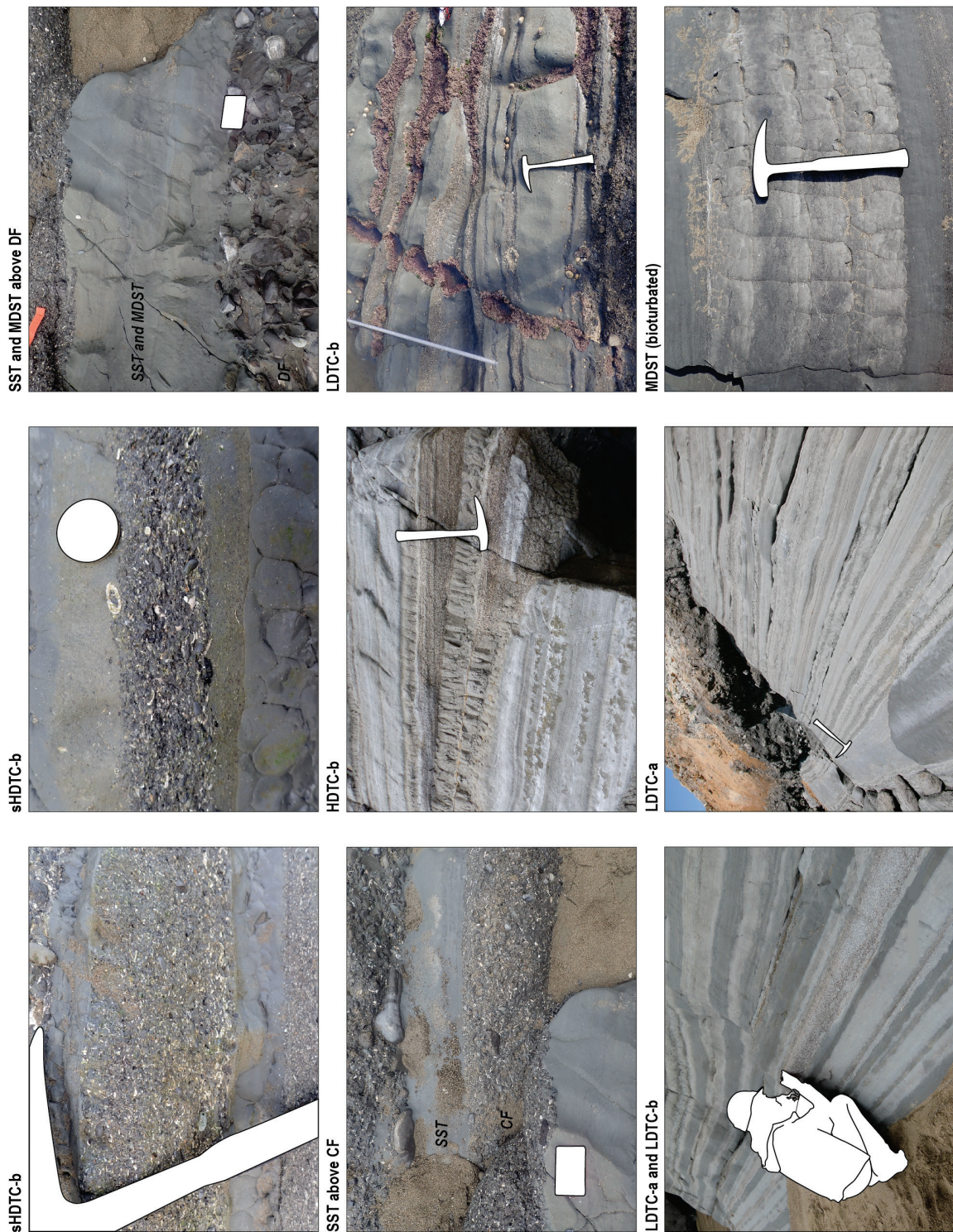
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Figure 6: Representative photographs of the lithofacies summarized in Table 1 and recognized at Homewood. Scales: human size (~1.70 meters); Jacob's staff (1.5 meter); hammer (33 centimeters); card (8.5 centimeters long); coin (2.3 centimeters).

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Figure 7: Representative photographs of the lithofacies summarized in Table 1 and recognized at Homewood. Scales: human size (~1.70 meters); Jacob's staff (1.5 meter); hammer (33 centimeters); card (8.5 centimeters long); coin (2.3 centimeters).

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They were then grouped into three main facies associations, which are related to main depositional systems (**Fa1g**, **Fa2c**, **Fa3p**) (Figure 8; Figure 9). The nomenclature we adopted follows the one defined by Bailleul et al. (2007), Bailleul et al. (2013) and Claussmann et al. (2021), in which **Fa1g** refers to

252 depositional lobe settings, **Fa2c** to confined slope channel settings and **Fa3p** to shelf-derived mass-
253 wasting systems.

254 Each facies association is described and interpreted in sequence below, directly in context of the outcrop
255 where they were recorded. Figure 10, Figure 11 and Figure 12 illustrate the geometric relationships that
256 exist between the different lithofacies and their associations as well as some of the sedimentary
257 structures at different scales. The nature of the extra- and intraformational clasts reworked in the
258 conglomeratic and coarse sandstone intervals is described separately (see Section 6) and summarized
259 in Figure 13 and Figure 14.

260 **5.2. Confined slope system**

261 The Waikaraka section crops out five kilometers south of the Uruti Point and extends across one kilometer
262 along the coast (Figure 3). Deposition occurred in the Middle Miocene, Lillburnian to Waiauian (Late
263 Langhian to Serravallian, 15.1 – 11.63 Ma, Appendix 2) and features the **Fa2c** depositional system, which
264 can be divided into two main depositional units (**Fa2c-e**, **Fa2c-l**), hereafter described and interpreted in
265 sequence. These units are bounded by major erosion surfaces. These prominent surfaces can laterally
266 be traced over several hundreds of meters (>600 meters), exhibit a relief of up to two meters, and
267 characteristically separate contrasting depositional facies (Figure 8).

268 **5.2.1. Confined slope channel: axis to off-axis, early-stage fills (Fa2c-e)**

269 Observations

270 **Fa2c-e** is characterized by crude fining- and thinning-upward series, dominated by coarse-grained facies.
271 Here, this association crops out over more than 600 meters laterally and is generally between 10 to 50
272 meters in thickness (Figure 8).

273 **Fa2c-e** mostly comprises amalgamated, very thick-bedded (one to >two meters), disorganized (**CF**) to
274 organized (**gHDTC**), polymict and clast-supported conglomerates (Figure 4; Figure 10; Figure 11; Table
275 1). The conglomerates commonly form continuous, broadly lenticular and fairly symmetrical bodies with
276 low-relief erosional bases, thereby indicating a cross-sectional, rather than longitudinal, view of them.
277 Their upper topographies are uneven and they are occasionally overlain by thin- to thick-bedded,
278 structured or structureless sandstones (**SST**), themselves often incised by the overlying conglomerates,
279 thus forming composite conglomeratic bodies (Figure 4; Figure 8; Figure 11; Table 1). The conglomerates
280 essentially evolve from (1) being internally disorganized (**CF**), only comprising local internal stratifications,

281 marked by subtle variations in clast grade and alignment, or in matrix concentration (**DF**) to (2) being
282 internally organized, displaying inverse- to normally- or only normally-graded patterns (**gHDTC**) (Figure
283 4; Figure 8; Figure 11; Table 1).

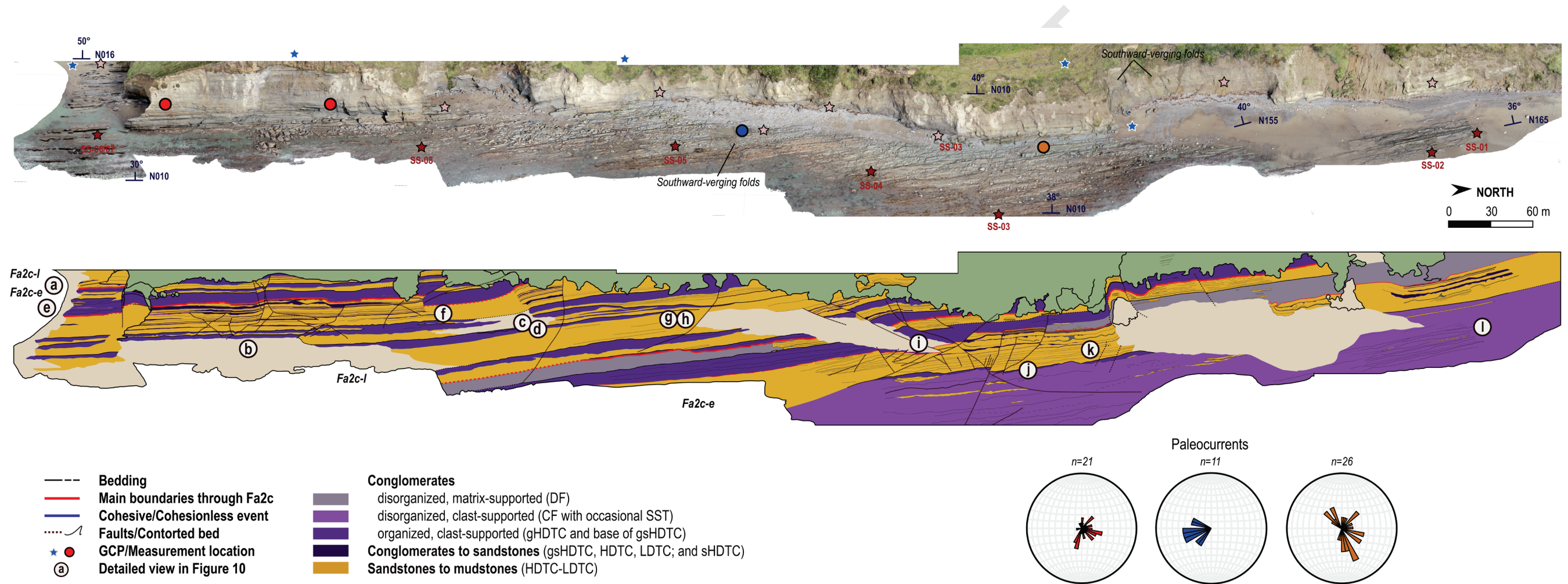
284 The conglomerates are generally overlain by series of laterally continuous, mostly medium- to fine-
285 grained, thin- to very thick-bedded sandstones, which are either amalgamated or capped by bioturbated
286 mudstones (**HDTC to LDTC**) (Figure 5; Figure 8; Figure 11j; Table 1). The thicker beds show frequent
287 grain-size breaks, indicating bed amalgamation (**HDTC to LDTC-b**). Their basal, coarser intervals are
288 discontinuous, erosive and typically composed of gravel material similar to that of the conglomerates,
289 which occasionally includes aligned mud clasts, plant debris and some shell fragments (Figure 5, LDTC-
290 b). The sandstones frequently contain soft-sediment deformation features (e.g., flame, load) in their
291 structured intervals (Figure 10k), whereas, the mudstone caps, where present, commonly include
292 *Phycosiphon* sp. and *Chondrites* sp. bioturbations.

293 Four occurrences of **Fa2c-e** were observed at Waikaraka (Figure 8; Figure 11).

294 The first occurrence is exposed at the base of the outcrop. Its basal surface is not visible as it lies below
295 the sea level. The conglomerates are disorganized (**CF**) (Figure 8; Figure 4; Figure 11) with an average
296 clast size ranging from granule- and pebble-grades, to pebble- and cobble-grades toward the southward
297 wavecut platform. Boulders and outsized clasts (deci- to decametric) are frequent (Figure 10l), particularly
298 to the south and in the disorganized intervals (**CF**). Sub-rounded clasts dominate, but, sub-angular clasts
299 are common and sometimes restricted to specific intervals (Figure 4; see Section 6). The overlying
300 sandstone-rich series generally fine and thin-upward, evolving from thick- to medium-bedded, concave-
301 up sandstones pinching out to the north into nearly-flat bottomed, thinner-bedded sandstones with very
302 thin- to thin-bedded mudstone caps (**HDTC to LDTC**) (Figure 8; Figure 10j, k; Figure 11). To the south,
303 amalgamation dominates throughout (Figure 10i; Figure 11). This first occurrence ends with a couple of
304 laterally continuous, very thick- to medium-bedded sandstones (**LDTC-b**) that are overlain by few thinner-
305 bedded intervals (**LDTC-a**) and characteristically comprise large-scale fluid escape and soft-sediment
306 deformation features (Figure 10j, k).

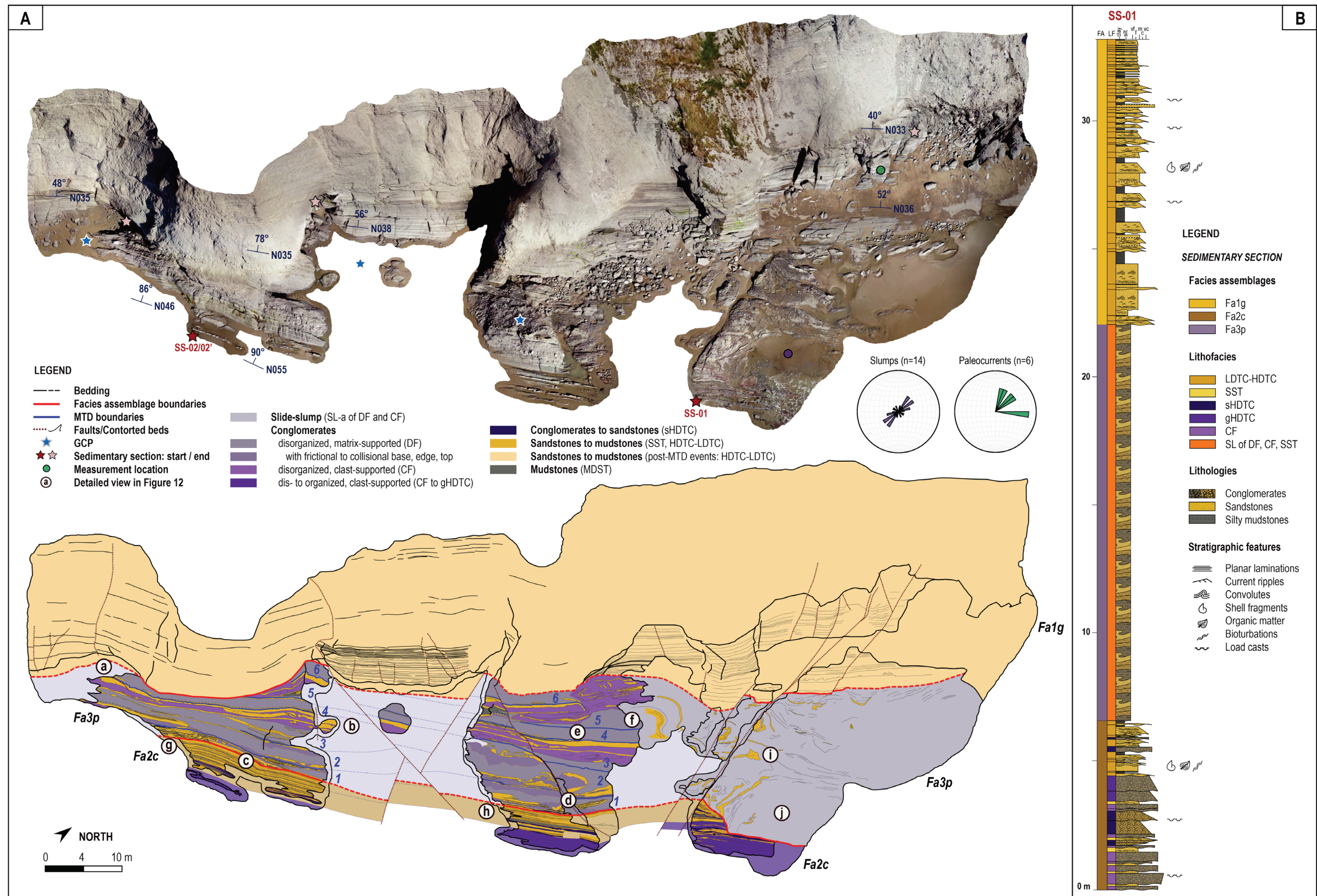
307 The second occurrence directly overlies the first, separated by a basal surface that varies from being
308 sharp to erosional (> two meters incision) to the south (Figure 8; Figure 10i; Figure 11). The associated
309 conglomerates are organized (**gHDTC**), mostly showing normally-graded patterns, amalgamated (i.e., no
310 capping sandstone) and smaller (~one to one and a half meters thick) (Figure 8; Figure 4; Figure 11).

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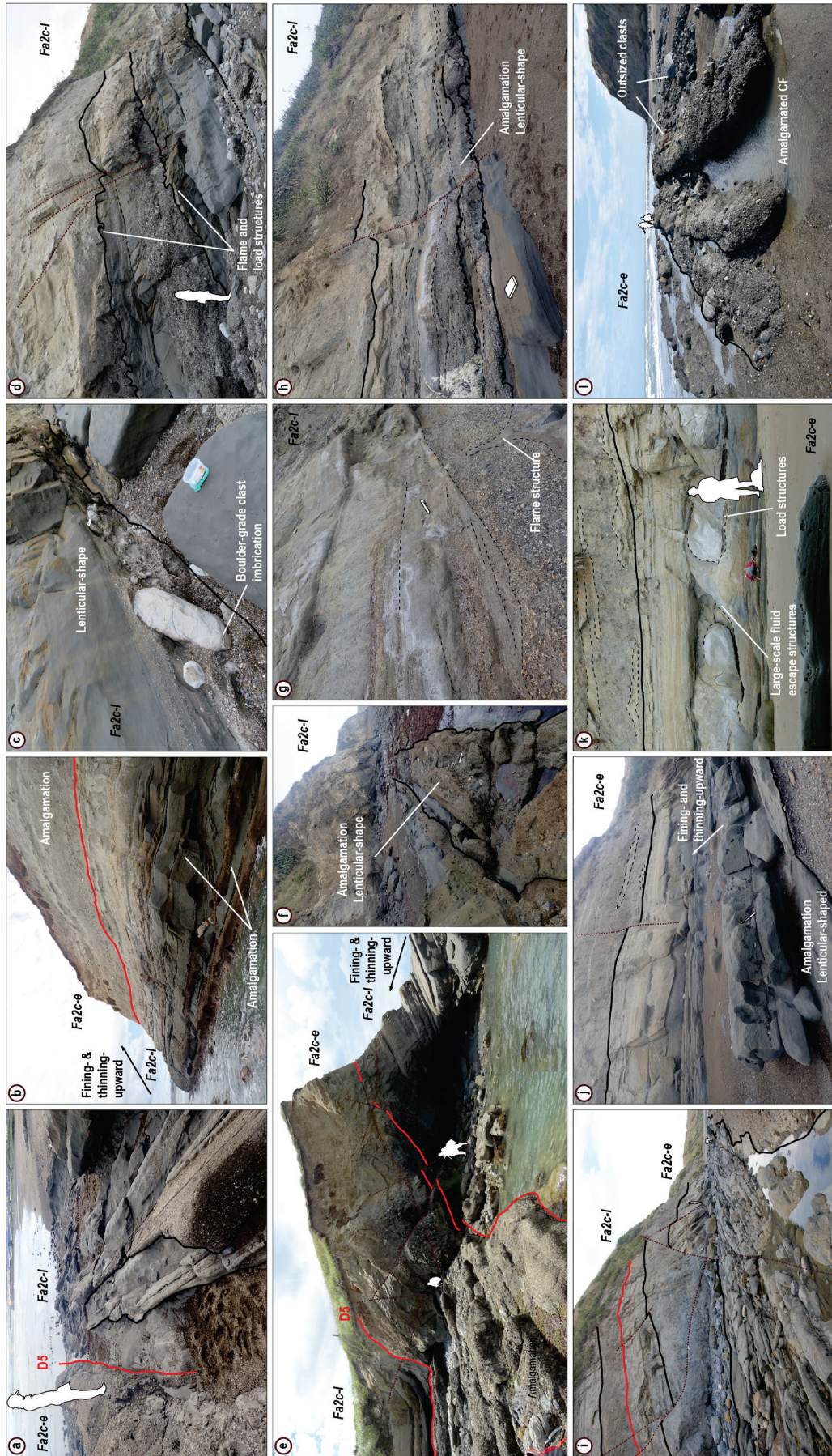
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317 Figure 8: 3D outcrop model (top) and interpretation (bottom) of the Waikaraka coastal outcrop that exposes a confined slope channel system (Fa2c). This system comprises two distinct channel complexes, essentially divided into early-stage confined channel fills (Fa2c-e) and late-stage
318 confined channel fills (Fa2c-l). The letters refer to remarkable architectural elements that characterize the Waikaraka system: a detailed view and explanation of each of them is available in Figure 10. The colors used in the stereoplots correspond to the location circles presented on the
319 outcrop model (top). Stereonets (Schmidt, lower hemisphere) highlight the paleocurrents taken in the turbidite system after back-tilting of bedding planes to initial horizontal position (assuming cylindrical folding).



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321 Figure 9: (A): 3D outcrop model (top) and interpretation (bottom) of the Homewood coastal outcrop. This outcrop exposes three distinct gravity-driven systems, from the bottom to the top, a slope channel system (Fa2c), a shelf-derived mass-wasting system (Fa3p) and a distal depositional
 322 lobe system (Fa1g). The letters refer to remarkable architectural elements that characterize the Homewood system: a detailed view and explanation of each of them is available in Figure 12. The colors used in the stereoplots correspond to the location circles presented in the outcrop model
 323 (top). Stereoplots (Schmidt, lower hemisphere) highlight the paleocurrents taken in the turbidite system (Fa1g, in green) and the fold axis of the slump-related folds taken in the mass-wasting system (Fa3p, in purple) after back-tilting of bedding planes to initial horizontal position (assuming
 324 cylindrical folding). The blue numbers U1 to U6 highlights six units, corresponding to six distinct episodes of mass wasting. (B): Sedimentary section 01 (SS-01) recorded at Homewood. This section provides an overview of the three facies associations encountered.



326 Figure 10: [previous page] Detailed views of the shelf-derived gravity-driven deposit main architectural and sedimentary
327 elements supporting the interpretation of the Waikaraka 3D outcrop model (Figure 8). Fa2c corresponds to the confined slope
328 channel system facies association, whereby Fa2c-e means early-stage fills and Fa2c-l late-stage fills. CF, DF, gHDTC,
329 gsHDTC, SST, HDTC, LDTC correspond to the lithofacies summarized in Table 1. (a): Discontinuity *D5* from [Bailleul et al.](#)
330 [\(2007\)](#) and [Bailleul et al. \(2013\)](#) marking the transition from Fa2c-e to the uppermost sand-rich deposits; (b, e): Fa2c-l fining-
331 and thinning-upward conglomerate (gHDTC, gsHDTC) to sandstone series (HDTC, LDTC) overlain by Fa2c-e amalgamated
332 conglomerates (CF and gHDTC) then followed by sandstones (HDTC, LDTC); (c): Fa2c-l boulder-grade clast imbrication
333 (base of gsHDTC); (d): Fa2c-l three successive channel storey fills characterized by conglomerate-rich bases (gsHDTC) with
334 upward (*e.g.*, flame) and downward (*e.g.*, load) soft-sediment features, overlain by sandstone-rich lithofacies (HDTC to
335 LDTC); (f, h): Fa2c-l amalgamated, broadly lenticular channel storey fills (gsHDTC, HDTC, LDTC); (g): Fa2c-l upward (*e.g.*,
336 flame) soft-sediment features and trough cross stratifications; (i): Fa2c-e transitioning to Fa2c-l; (j): Fa2c-e amalgamated
337 sandstones (HDTC) pinching out to the north overlain by the Fa2c-e conglomerates (DF, CF); (k): large-scale fluid escape
338 structures and downward (*e.g.*, load) soft-sediment features in the Fa2c-e sandstones series (HDTC, LDTC) that underlies
339 the next Fa2c-e conglomerate events (DF, CF); (l): Fa2c-e amalgamated conglomerates with outsized clasts (CF).

340 The deposits are moderately-sorted and in average comprise sub-rounded granule- to pebble-grade
341 clasts with a few scattered boulders and outsized clasts (Figure 4; see Section 6). To the north, these
342 conglomerates are matrix-supported (Figure 8) and locally include a very thin, clast-supported basal layer
343 (**sheared DF**) (Figure 4). The overlying sandstones characteristically present well-developed, southward-
344 oriented soft-sediment deformation features in their structured intervals and their beds are frequently
345 contorted, presenting apparent southward-verging recumbent folds (Figure 8). To the south, these folds
346 appear to eventually evolve into matrix-supported conglomerates containing both folded sandstones and
347 extraformational pebble-grade clasts (**SL-a to DF**) (Figure 8; Figure 11).

348 Finally, the third occurrence is present toward the top of the Waikaraka outcrop and alternatively
349 comprises disorganized (**CF**) and organized (**gHDTC**) conglomerates, sometimes overlain by structured
350 to structureless sandstones (**SST**) (Figure 8; Figure 10e; Figure 11). The deposits are generally
351 moderately sorted and, on average, comprise sub-rounded pebble- to cobble-grade clasts (Figure 4; see
352 Section 6), although some occurrences carry boulders and outsized clasts (Figure 10e). The overlying
353 sandstone series is only a few meters thick, interrupted by another couple of meter-thick conglomeratic
354 episode (fourth occurrence), itself characterized by a sharp upper surface (Figure 10a).

355 Interpretations

356 **Fa2c-e** is interpreted to characterize the axis to off-axis, early-stage (Stage I of [Kneller et al. \(2020\)](#)) fills
357 of confined slope channel complexes (*sensu* [Sprague et al. 2005](#)).

358 The laterally extensive, coarse-grained and clast-supported conglomerates at its base represent
359 deposition from gravelly gravity flows (Lowe 1982, R1, R2, R3 divisions). Upslope failures may have
360 triggered all or part of these flows, which then began to transform from cohesive to cohesionless debris
361 flows downslope (Lowe 1982; Postma 1986; Sohn et al. 2002). Their abundant amalgamation surfaces,
362 clast-supported textures (e.g., disorganized, inverse, normal) and very coarse grain sizes indicate
363 significant bed load transport and rapid deposition of the gravels underneath gravelly high-density
364 turbidity currents in an environment dominated by high-energy flows, erosion and bypass (Lowe 1982).
365 The abundant scoured bases, combined with the other textural and sedimentary features suggest that
366 these flows were funneled through low sinuosity, braid-like channels, as recurrently found in base of
367 confined slope channel complexes (e.g., Galloway 1998; Di Celma et al. 2010; Gamberi et al. 2013; Li et
368 al. 2018; McArthur and McCaffrey 2019; Kneller et al. 2020). The southward increase in clast size
369 observed in the first occurrence suggests a southward evolution from off-axis to axis settings (Campion
370 et al. 2000).

371 The general upward internal organization observed in the conglomerates and the overlying sandstone
372 series likely resulted from gravelly to sandy high- to low-density turbidity currents from multiple, rather
373 than single flows (Lowe 1982), with flows initiating from a shallow marine environment connected to a
374 vegetated hinterland sourcing the abundant terrigenous material (Kuenen 1964). The resulting deposits
375 suggest lower energy environments leading to both the progressive abandonment of the channel system
376 and passive infilling of the relief (Galloway 1998; McHargue et al. 2011). Axis to off-axis abandonment
377 facies are typically sand-rich, fairly thick (e.g., medium-bedded) and highly to moderately amalgamated
378 (Campion et al. 2000; McHargue et al. 2011; Hubbard et al. 2014).

379 As previously described, four distinct occurrences of these axis to off-axis, early-stage fills were observed
380 at Waikaraka, thereby indicating the presence of several slope channel complexes infilling one larger
381 channel conduit (*sensu* Sprague et al. 2005). However, **Fa2c-e** only provides information on the deposits
382 that form the lower part of these channel complexes.

383 The textural changes observed in the second occurrence here suggest transformation from cohesive to
384 cohesionless flows, resulting from progressive dilution of a parental debris flow (DF) evolving into a high-
385 concentration gravelly dispersion overlain by a lower-concentrated and turbulent suspension (gHDTc)
386 (Sohn et al. 2002, *i.e.*, bipartite or stratified flows). When taken in consideration with the recurrent soft-
387 sediment deformations, these elements suggest that this second occurrence may originate from several,
388 small-scale failure events, either at or near the shelf break, or directly from the channel walls (Di Celma
389 et al. 2010; Janocko et al. 2013; Janocko and Basilici 2021). The resulting mass-wasting products

390 deposited in the channel conduit and the southward component observed in the above features indicate
391 a potential northward location of the source region. Whether it is regional or local is here a matter of
392 speculation. The large-scale fluid escape structures present in the underlying sandstones would result
393 from sudden loading of the mass-wasting products onto the sandstone strata, increasing pore pressure
394 and promoting dewatering (Figure 10k) (Browne et al. 2020).

395 Despite the changes observed in the second occurrence's texture and clasts (see Section 6.3), the
396 absence of shutdown mud drape, the only weakly erosional basal surface separating the first and second
397 occurrences, and their general crude fining- and thinning-upward trend all suggest that these two
398 occurrences belonged to the same channel complex, rather than two distinct complexes (Kneller et al.
399 2020). The same reasoning applied to the third and fourth occurrences allows considering that these two
400 belonged to another confined channel complex.

401 Overall, the lateral and vertical facies evolution observed throughout **Fa2c-e** suggests a general evolution
402 from off-axis to axis settings towards the south. Notwithstanding the overall outcrop conditions (no 3D
403 control) and or the orientation of the outcrop to withhold their exposures, the absence of recognizable
404 lateral-accretion packages (LAPs) (*sensu* Abreu et al. 2003) may also inform that the slope channel
405 system (**Fa2c**) was likely low to moderately sinuous and generally erosional. Such channel belts usually
406 occur in the upper- to middle-parts of the continental slope (Janocko et al. 2013). Finally, the
407 micropaleontological analyses suggest that deposition took place at outer bathyal paleodepths (Appendix
408 2).

409 **5.2.2. Confined slope channel: axis to off-axis, late-stage fills (Fa2c-l)**

410 Observations

411 **Fa2c-l** presents contrasting sedimentary facies and architectural styles to those of the underlying **Fa2c-**
412 **e**, characteristically comprising fining- and thinning-upward series, dominated by fine-grained facies, only
413 punctuated by coarser intervals. This association is traceable across the entire outcrop laterally (>900
414 meters) and can hold over 40 meters in thickness.

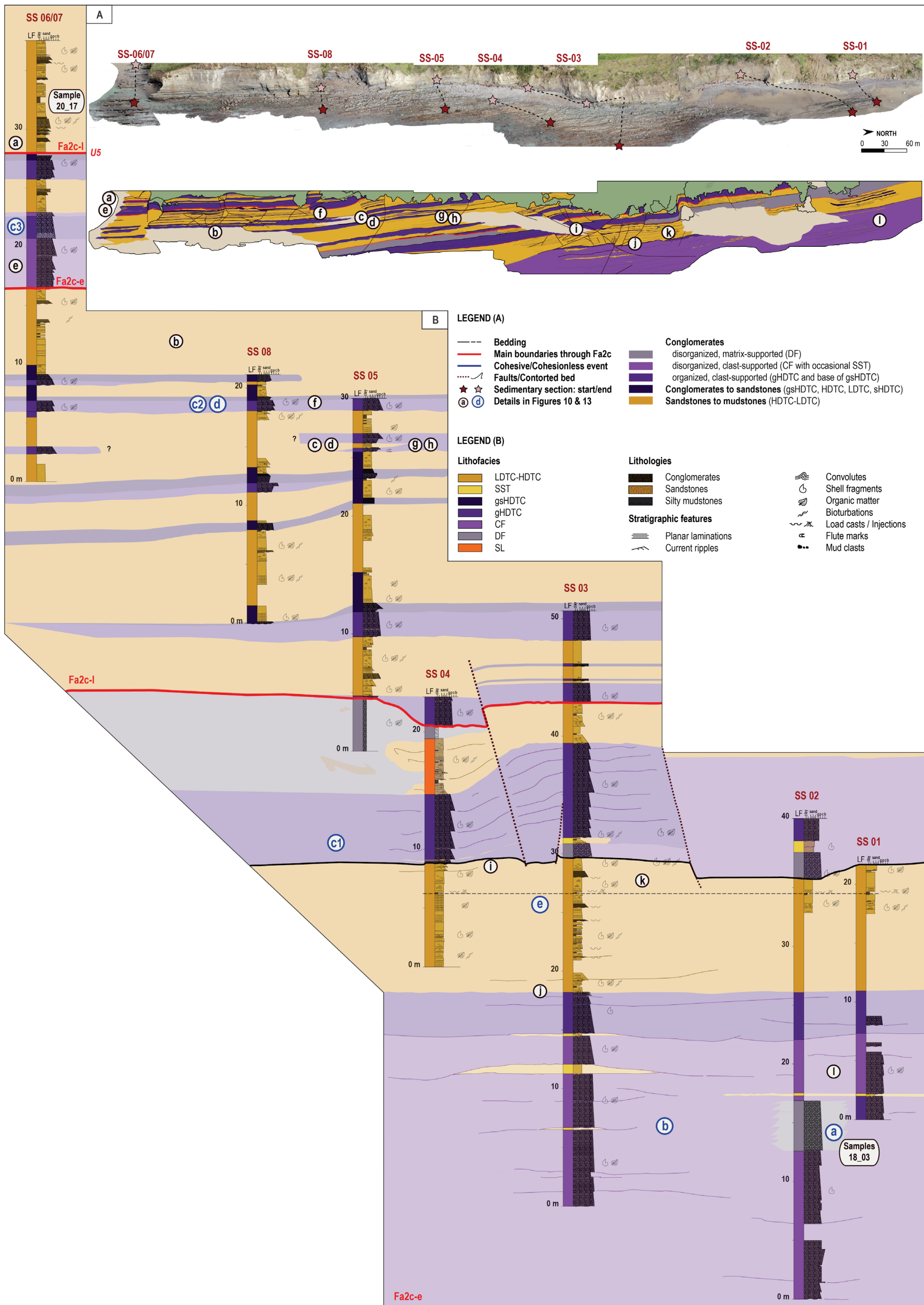


Figure 11: Sedimentary sections recorded at the Waikaraka outcrop. This system comprises two distinct channel complexes, respectively divided into the early-stage confined channel fills (Fa2c-e) and late-stage confined channel fills (Fa2c-l). The letters in dark red refer to remarkable architectural elements that characterize the Waikaraka system: a detailed view and explanation of each of them is available in Figure 10. The letters and numbers in blue give the location of the clast analyses summarized in Figure 13. Details on the sample data are available in Appendix 2. The red surface named D5 at the top of the Waikaraka fill corresponds to the discontinuity D5 from (Bailleul et al. (2007) and Bailleul et al. (2013), which marks the transition to the period of generalized subsidence that affected the Hikurangi Margin during the Middle Miocene, Waiuan.

1 **Fa2c-I** comprises eight successive channel storey fills (*sensu* Sprague et al. 2005) (Figure 8; Figure 11).
2 Each storey commonly starts with amalgamated, thick- to very thick-bedded (one to one and a half
3 meters), organized, polymict and clast-supported conglomerates (**gHDTC**) deposited above a broadly
4 erosional, concave-up base (Figure 4; Figure 11). These are overlain by a fining- and thinning-upward
5 conglomerate to sandstone series, capped by bioturbated mudstones (**gsHDTC**), which is, in turn,
6 overlain by another fining- and thinning-upward series of nearly-flat bottomed, thick- to thin-bedded
7 sandstones and mudstones (**HDTC** to **LDTC**) (Figure 5; Figure 10b, h; Figure 11). The basal
8 conglomerate to sandstone series shows repeated bed amalgamation, thus forming composite
9 stratigraphic infills (Figure 10f; Figure 11). Conglomerates are concentrated within the basal intervals;
10 sandstones quickly become dominant, generally representing a third of the channel storey fill (Figure 8;
11 Figure 11). The conglomeratic intervals are broadly lenticular, laterally pinching out, amalgamated and
12 often include clast imbrications and subtle cross-stratifications (Figure 10c, f, h). Interestingly, their matrix
13 seems primarily made of bioclast content (*e.g.*, shell fragments) toward the top of **Fa2c-I**.

14 The bases of the conglomerates commonly display centi- to decametric soft-sediment deformation
15 structures (*e.g.*, flames) and large sole marks (*e.g.*, flutes) (Figure 5; Figure 10d, g). Boulders and
16 oversized clasts usually mark the transition from the inverse- to normally-graded patterns, and their
17 proportion generally increases toward the south (Figure 5, **gsHDTC** to **HDTC**). Clasts are sometimes of
18 larger size than the conglomerate bed itself, thus protruding into the overlying sandstones (Figure 5,
19 **gsHDTC** to **HDTC**). On average, sub-rounded pebble-grade clasts dominate. In **Fa2c-I**, the **HDTC**
20 sandstones that directly overlie the conglomerates are commonly coarse-grained in their traction
21 intervals, show trough cross-stratifications and include plant debris, numerous shell fragments (Figure
22 10g) and abundant aligned mud clasts at their base. These sandstones form broadly lenticular bodies,
23 with locally erosive bases characterized by very coarse- to coarse-grained material (Figure 10a, f). They
24 eventually reach medium- to fine-grain sizes in the remaining intervals and in the subsequent fining- and
25 thinning-upward series (**HDTC** to **LDTC**). The paleocurrents measured in the sandstone intervals indicate
26 a general southward trend (Figure 8), which is consistent with measurements from imbrications within
27 the coarse-grained basal fills.

28 Two occurrences of **Fa2c-I** were observed at Waikaraka (Figure 8; Figure 11).

29 The first occurrence corresponds to the above description. It is about 40 meters thick and comprises eight
30 channel storeys (*sensu* Sprague et al. 2005). Its basal surface is sharp, concave-up and broadly incises
31 **Fa2c-e** with up to 2 meters relief (Figure 8). The second occurrence is observed at the top of the
32 Waikaraka outcrop and only crops out for about 15 meters laterally and vertically, being covered by

33 quaternary beach deposits after that. Its basal surface is sharp above the underlying **Fa2c-e** occurrence
34 and it is only comprised of sandstones (**HDTC to LDTC**) (Figure 10a, e; Figure 11).

35 Interpretations

36 **Fa2c-l** is interpreted to represent the axis to off-axis, late-stage (Stage II of Kneller et al. (2020)) fills of
37 confined slope channel complexes (*sensu* Sprague et al. 2005). Characterized by multiple episodes of
38 erosional bypass and sedimentation, these form the middle fills of channel complexes.

39 The conglomerate- to sandstone-dominated basal successions (**gHDTC** and **gsHDTC**) that broadly fill
40 the erosion surface result from stratified gravelly gravity flows that are confined within channels (Lowe
41 1982; Campion et al. 2000; Sohn et al. 2002). The basal dewatering features suggest rapid deposition
42 and loading of the channel fills, whereas the frequent amalgamation observed in both the conglomerates
43 and sandstones implies that erosion and bypass were dominant in the basal interval (Lowe 1982).
44 Transport of large (*e.g.*, oversized) clasts can be explained by flow confinements such as the ones found
45 in confined slope channels, where the flow size, velocity and carrying capacity are maximized (Postma
46 et al. 1988). The high degree of preservation of mud clasts in both the conglomerates and overlying
47 sandstones, scattered as lag material indicates high fall-out rates (Postma 1986), whereas the increase
48 in content of coarse and fine skeletal fraction toward the top of the series rather suggest deposition of
49 calciturbidites (*e.g.*, Haak and Schlager 1989; Reijmer et al. 2015). The gravelly gravity flows then
50 decelerated and became less confined. The overlying tabular, fining- and thinning-upward series of
51 sandstones and mudstones (**HDTC to LDTC**) mark the transition to lower energy depositional
52 environments, resulting in passive filling and progressive abandonment of the channel system (Galloway
53 1998; McHargue et al. 2011). As in **Fa2c-e**, such sand-rich abandonment facies here suggest axis to off-
54 axis settings (Campion et al. 2000). Overall, these broadly lenticular and laterally extensive channel
55 storey fills are dominated by sandy high- to low-density turbidites, scouring a mud-rich substrate during
56 downslope transport and or at hydraulic jumps (*i.e.*, abundant mud clasts) (Fonnesu et al. 2016). The
57 vertically aggrading, semi-amalgamated style of the channel storey fills confirms a generally confined
58 slope channel complex settings (Sprague et al. 2005), but less steep and confined to those of the
59 previously described **Fa2c-e**.

60 Although most likely belonging to the same slope channel system, the uppermost occurrence
61 characteristics and micropaleontological analyses both suggest contrasting depositional settings. Not
62 only the micropaleontological study indicates lower paleodepths of deposition (*i.e.*, lower bathyal), it also
63 indicates that this latter series was restricted to the Middle Miocene, Waiauian (Serravallian, 13.05 – 11.63
64 Ma) (Figure 11; Appendix 2).

65 Overall, we interpret the Waikaraka outcrop to represent the axis to off-axis stratigraphic infills of two
66 distinct channel complexes, essentially capturing the early- to late-stage fills of a larger, yet confined
67 slope channel system (*sensu* Sprague et al. 2005). We further discuss the controls on such a system and
68 the implications of its uppermost deposits in Section 7.

69 **5.3. Weakly confined slope and depositional lobe systems**

70 The Homewood section crops out one kilometer north of the Kaiwhata river mouth and extends over 110
71 meters along the coast (Figure 3). It features three distinct Middle Miocene, Lillburnian (Late Langhian to
72 Serravallian, 15.1 – 13.05 Ma, Appendix 2) gravity-driven systems described and interpreted below as
73 three separate facies associations (**Fa2c**, **Fa3p**, **Fa1g**) (Figure 9).

74 **5.3.1. Slope channel (Fa2c)**

75 Observations

76 At Homewood, **Fa2c** is characterized by crude fining- and thinning-upward series, evolving from coarse-
77 to fine-grained facies. Owing to outcrop conditions, it is intermittently exposed for about 80 meters
78 laterally and only represents a ten of meter-thick succession of sub-vertical beds. Its basal surface does
79 not crop out and it is overlain by the erosive **Fa3p** (Figure 9).

80 Here, **Fa2c** is characterized by a series of tabular, continuous (1) medium- to thick-bedded (10 to 50
81 centimeters), disorganized (**CF**) to organized (**gHDTC**) clast-supported conglomerates with sandstone
82 caps (**SST**), followed by (2) thin- to medium-bedded (3 to 15 centimeters) pebbly sandstones (**sHDTC-a**)
83 or bioclastic grits (**sHDTC-b**), either interbedded with mudstones (**MDST**) or structureless sandstones
84 (**SST-b**) and mudstones (**MDST**), finally overlain by (3) medium- to thick-bedded (10 to 50
85 centimeters) sandstones with mudstone caps (**LDTC-b**) (Figure 6; Figure 7; Figure 9; Figure 12g, h).

86 Overall, a greater internal organization is observed in **Fa2c** to the north-east of the outcrop.
87 Conglomerates are generally ungraded (**CF**) to the south-west of the outcrop, whereas they gradually
88 show internal organization (**gHDTC**), displaying inverse- to normally- or only normally-graded patterns to
89 the north-east (Figure 6). The conglomerates mostly consist of sub-angular to sub-rounded granule- to
90 pebble-grade clasts (see Section 6). Cobbles and boulders are locally observed. Similarly, the bioclastic
91 grits (**sHDTC-b**) become normally-graded, and include mud clasts and or bioclastic fractions aligned
92 along horizons towards the north-east of the outcrop (Figure 7). In both cases, the basal surfaces of the
93 conglomerates and bioclastic grits are slightly erosive and commonly display load structures.

94 Finally, the sandstones (**LDTC-b**) to the south-west present rare, discontinuous, centimeter-thick, very
95 coarse basal intervals (Ta), whereas to the north-east, these basal intervals are well-developed, fairly
96 continuous and include frequent mud clasts and bioclastic fractions aligned along horizons, as well as
97 load structures (Figure 7). The related sandstones also commonly show distinct grain-size breaks,
98 indicating amalgamation of beds.

99 Interpretations

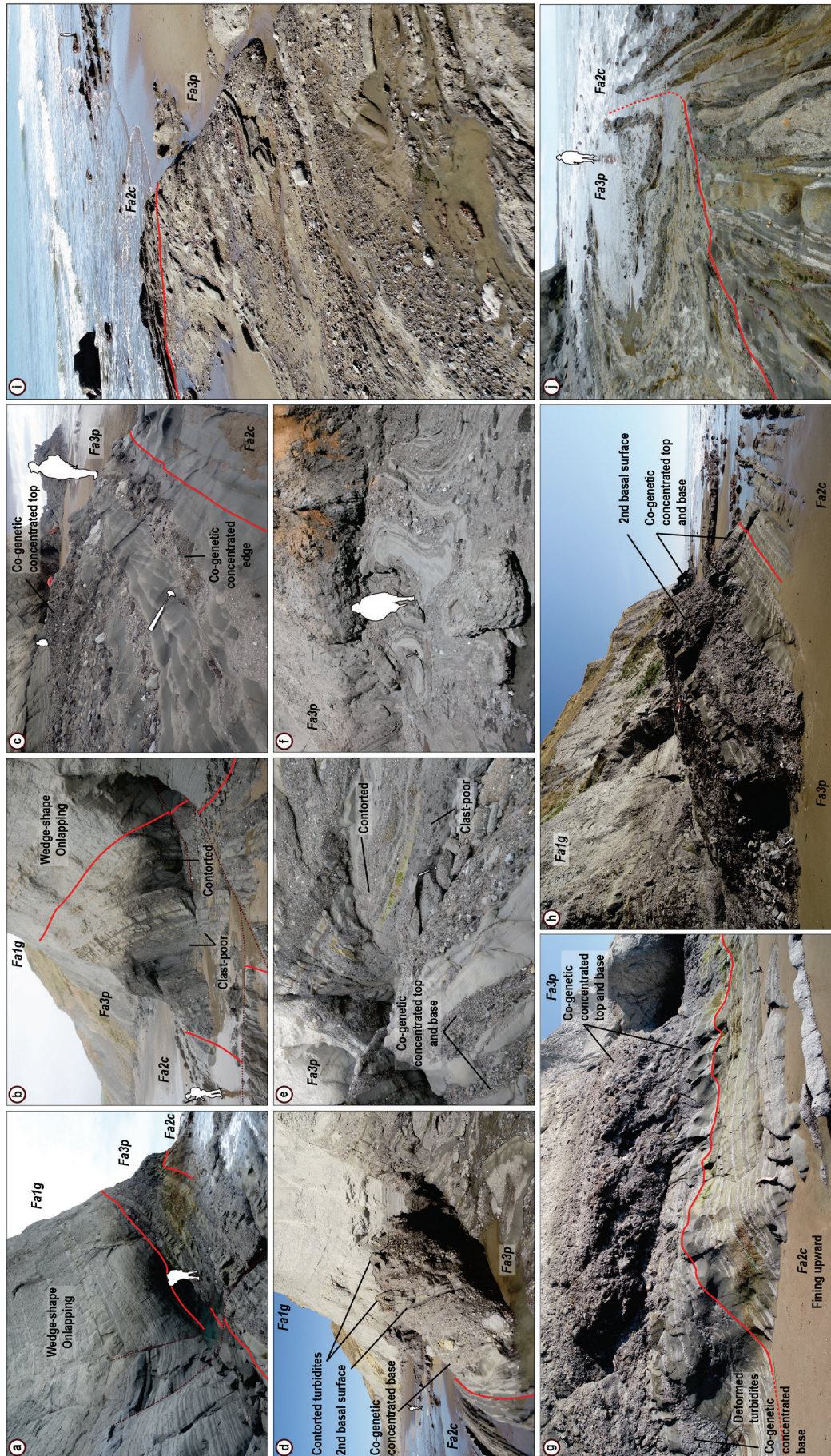
100 We interpret this basal succession to result from both flow processes and depositional environments
101 (**Fa2c**) similar to those described at Waikaraka. While outcrop exposure hinders recognition of specific
102 architectural elements, such as channels, and of particular depositional units (**Fa2c-e**, **Fa2c-l**), the
103 stratigraphic infill outlines comparable crude fining- and thinning-upward conglomerate to sandstone
104 series, resulting from gravelly to sandy high- to low-density turbidity currents (Lowe 1982). The main
105 difference resides in the scale of the individual beds, being significantly smaller at Homewood, which
106 could either point towards (1) smaller source region or slope-channel system, and or (2) contrasting
107 locations along the slope or laterally within the channel.

108 **5.3.2. Mixed shelf-derived mass-transport deposits (Fa3p)**

109 Observations

110 **Fa3p** mostly comprises interbedded coarse-grained facies, here exposed across more than 120 meters
111 laterally and about 15 meters thick (Figure 9).

112 **Fa3p** is characterized by a succession of thick- to very thick-bedded (<one to >two meters), disorganized,
113 polymict, matrix- (**DF**) and clast-supported (**CF**) conglomerates, that laterally becomes entirely deformed
114 and contorted (**SL-a**) to the north-east of the outcrop (Table 1; Figure 6; Figure 7; Figure 9). The
115 conglomerates are recurrently capped by thin- to thick-bedded, either structureless or structured
116 sandstones (**SST**) (Table 1; Figure 6; Figure 7; Figure 9; Figure 12), the latter of which can sometimes
117 evolve to a few interbedded structured, very thin- to medium-bedded sandstones with mudstone caps,
118 commonly bioturbated (**LDTC-a**) (Table 1; Figure 7; Figure 9; Figure 12b, e).



120 Figure 12: [previous page] Detailed views of the shelf-derived gravity-driven deposit main architectural and sedimentary
121 elements supporting the interpretation of the Homewood 3D outcrop model (Figure 9). Fa2c corresponds to the confined
122 slope channel system facies association, Fa3p to the shelf-derived mass-wasting system and Fa1g to the distal depositional
123 lobe system. SL, CF, DF, gHDTc, SST, sHDTc, HDTc, LDTc correspond to the lithofacies summarized in Table 1. (a, b):
124 Fa3p succession of interbedded conglomerates (DF, CF) recurrently capped by sandstones (SST) and overlain by the
125 onlapping Fa1g sandstones (HDTc, LDTc); (b): Fa3p varying bed dips and upper contorted unit; (c, d, e, g, h): Fa3p
126 conglomerates showing local variation in clast concentrations at their base, edge and or top (sheared DF), (d, h): Fa3p sharp
127 surface at the base of the succeeding conglomerate, which holds contorted sandstones; (e): Fa3p occurrence of clast-poor
128 conglomerate (DF) and upper contorted unit; (f, i, j): Fa3p contorted conglomerates (SL-a, made of DF, CF and SST); (g, h):
129 Fa2c crude fining- and thinning-upward conglomerate to sandstones (gHDTc, sHDTc, LDTc), overlain by Fa3p, which is
130 characterized by erosion as well as local substrate entrainment and deformation at its base; (i, j): Fa3p basal eroded through
131 Fa2c towards the north-east.

132 The basal surface is highly erosive (>8 meters) towards the north-east (Figure 9, red surface at the base
133 of Fa3p; Figure 12g, i, j), and both local entrainment and deformation (*i.e.*, apparent northward-verging
134 contorted sandstones) of the substrate are observed (Figure 9; Figure 12g). In contrast, the upper surface
135 is highly irregular, with overlying beds onlapping on a deci- to decametric scale (Figure 9; Figure 12a, b).
136 Laterally, the upper surface can appear rather flat and overlain by relatively tabular sandstones. The first
137 overlying sandstone beds are generally internally chaotic; they sometimes incorporate material from **Fa3p**
138 (*e.g.*, pebbles and cobbles) and can form well-developed, decametric-scale soft-sedimentation
139 deformation structures.

140 Five additional internal surfaces were also identified, which allows division of **Fa3p** into six units (Figure
141 9, blue surfaces within Fa3p). Commonly observed at the top of the structureless or structured beds
142 (**SST**), these surfaces are sharp and characterized by thin- to medium-bedded silty mudstones (**MDST**)
143 that can be traced across the outcrop (Figure 9, blue surfaces; Figure 12d, h).

144 In the southern and central parts of the outcrop, **Fa3p** comprises fairly tabular beds. Their thicknesses
145 tend to increase north-eastward, thus generally forming lens-like shapes opening towards the north-east
146 (Figure 9). The beds' orientation changes from N055 to N035 and the dip measurements show an abrupt
147 evolution from 90° (sub-vertical) to ~60° towards the west in only 15 meters. The dip of the first four units
148 (separated by the internal surfaces described above) changes from 90° to 78°, whereas the remaining
149 two directly onlap onto the preceding ones with a dip starting around 70° and evolving towards 60°. To
150 the north-east, **Fa3p** is completely disorganized and shows SW-NE oriented recumbent folds (Figure 9,
151 slump measurements; Figure 12j).

152 Three different scales of internal contortion are observed in **Fa3p**, evolving both vertically and laterally:
153 (1) at clast-scale; (2) at unit-scale and (3) at the scale of several units. First, the entrained sandstones

154 display both apparent NE-verging (Figure 9) and NE-SW-oriented recumbent folds (Figure 12f). Second,
155 the fifth unit, previously observed to onlap on the underlying beds, appears to be severely folded, with a
156 NE-SW main axis (Figure 9, unit 5; Figure 12b, e). Third, the upper **Fa3p** units become remobilized
157 laterally, folding following an apparent NE-verging trend (Figure 12f). To the north-east of the outcrop,
158 this remobilization involves the entire facies association (*i.e.*, interbedded **DF**, **CF** and **SST**) (Table 1;
159 Figure 9; Figure 6; Figure 12f, i, j).

160 The matrix of the interbedded polymict conglomerates generally represents 40 to 70% of the **DF** deposits
161 with two occurrences comprising up to 80% of matrix (Figure 6; Figure 12b, e; Figure 13), and <5 to 15%
162 in **CF**. **DF** can display slight upward increase in matrix content. Alternatively, **DF** commonly shows vertical
163 and or lateral variations in both clast concentration and size at its base, top and or edges that appear
164 akin to the clast-supported lithofacies (**CF**) (Figure 9; Figure 6; Figure 12c, d, e, g, h). These concentrated
165 intervals are characterized by clasts of smaller grades (*i.e.*, granule- to pebble-grade) than those of the
166 general matrix-supported intervals (*i.e.*, pebble- to cobble-grade) (Figure 6). Overall, the average clast
167 size is of pebble- to cobble-grade in **DF**, whereas granule- to pebble-grade dominates **CF**. In **DF**, larger
168 clasts are frequent and can be outsized (deci- to decametric); conversely, they are rare and of cobble-
169 grade in **CF**. In both **DF** and **CF**, clast edges typically vary between sub-angular to sub-rounded shapes.
170 The highest proportions of sub-angular clasts are found in the largest clasts (*e.g.*, boulders). Sub-rounded
171 to rounded clasts are mostly present in the concentrated granule- to pebble-grade intervals of **DF** or in
172 some rare occurrences of **CF**.

173 Interpretations

174 We interpret **Fa3p** to represent deposition (*i.e.*, mass-transport deposits [MTDs]) from successive mass-
175 wasting events that triggered recurrent debris flows (**DF** and **CF**) (Postma 1986; Mulder and Cochonat
176 1996). The clast-supported debris flows (**CF**) are inferred to result from cohesionless flows (Postma
177 1986). Even if matrix strength dominates in **DF**, thereby suggesting cohesive flows (Nardin et al. 1979;
178 Lowe 1982), Nemeč and Steel (1984) demonstrated that the predominance of matrix in conglomerates
179 was not always a criterion to discriminate cohesive and cohesionless debris flows. Although the
180 ungraded, disorganized and matrix-rich occurrences of **DF** result from cohesive flows (Johnson 1984;
181 Postma 1986), the 2-to-3 phase occurrences (**sheared DF**) show internal organization that indicates fully
182 sheared flows with additional support mechanisms, such as grain collisions and dispersive pressure,
183 characteristic of cohesionless debris flows (Lowe 1982; Nemeč and Steel 1984; Postma 1986).

184 Multiphase debris flows require steep slopes (Postma 1986). The lens-like shapes observed in some of
185 their deposits may suggest channelization, characteristic of steep slope settings (Ortiz-Karpf et al. 2017;

186 Calhoun and Clague 2018). The high-amplitude variations recorded in the dip of beds (~40°) indicates
187 that they were syn-kinematic with the rise of a structure to the east, *i.e.*, growth strata. Changes in slope
188 gradient chiefly impact flow behavior (*e.g.*, velocity) and may promote flow transformations (*e.g.*, dilution,
189 shearing) (Fisher 1983), which in turn control the nature and geometry of the resultant deposits. Hence,
190 slope angle changes could partly explain the variety of debris flow deposits recorded in **Fa3p**. Also, the
191 considerable amounts of granule- and pebble-grade clasts (Sohn et al. 2002) as well as the recurrent
192 presence of mudstone layers and fine sandstones (**SST**), respectively under- and overlying the
193 conglomerates, all point toward fast-moving debris flows that hydroplane (*sensu* Mohrig et al., 1998). The
194 multiphase debris flows described here would thus result from efficient dilution of cohesive, hydroplaning
195 debris flows (Mohrig et al., 1998); dilution which can arise only a few kilometers from the source areas
196 (Sohn et al. 2002) and which can lead to the generation of slower-moving suspended sediment clouds
197 eventually depositing above them (**SST**) (Mohrig et al. 1998; Sohn et al. 2002; Mohrig and Marr 2003;
198 Talling et al. 2004). The first debris flow, however strongly erosional, suggests both an abrupt change in
199 the depositional settings to that of the underlying **Fa2c** deposits and an initially different mechanism (*i.e.*,
200 substrate erosion) responsible for flow mobility (Sobiesiak et al. 2018).

201 The internal surfaces allow to distinguish at least six mass-wasting events. The different scales of internal
202 contortion and deformation observed in the **Fa3p** deposits however complicates their depositional history,
203 thereby questioning the hypothesis of simple deposition of one mass flow on top of another. Basal
204 deformation of the substrate linked to the passage of an overriding flow is common, however the flow
205 mobility mechanisms proposed here (*i.e.*, hydroplaning) diverges from such a scenario, promoting little
206 to no interaction between the moving flow and underlying deposits for five of the units (Sobiesiak et al.
207 2018). Notwithstanding the possibility for the contorted strata to shear and deform in the direction of the
208 flow and thus capture the paleoslope direction at the clast-scale (Fonnesu et al. 2016), we here suggest
209 a different mechanism to explain the deformation recorded at the unit-scale and propose that a change
210 in slope gradient can both trigger development of a mass flow upslope and coeval remobilization
211 (translation and deformation) of the previously deposited and only poorly lithified mass-flow deposits
212 downslope. When combined with bed dips, the analysis of the fold hinges and axial planes at both clast-
213 and unit-scales consistently indicates a NW-dipping paleoslope and transport direction (Alsop et al. 2019;
214 Alsop and Weinberger 2020), coherent with a rising structure to the east.

215 The larger-scale remobilization observed to the north-east that involves the entire facies association **Fa3p**
216 could also be associated with this growing structure. As such, we interpret it to result from local failure of
217 the entire series due to the recurrent loading of sediments onto this rising slope, two mechanisms known

218 to greatly affect slope stability and encourage its failure (Hein and Gorsline 1981). Finally, the
219 micropaleontological studies imply an upper- to mid-bathyal depositional environment (Appendix 2).

220 5.3.3. Distal depositional lobe system (Fa1g)

221 Observations

222 **Fa1g** is characterized by laterally continuous, on average medium-bedded (~14 centimeters) and fine-
223 grained sandstones that show decreasing bed dips upward (56° to 32°). This association is exposed
224 along ~one kilometer of coastline and is over 200 meters in thickness (Appendix 4). Only the lowermost
225 part of this system is described in context of the outcrop model presented in this study (Figure 9; Fa1g).

226 **Fa1g** mostly comprises interbedded tabular, medium- to thick-bedded, frequently amalgamated
227 sandstones (**LDTC-b**) and very thin- to medium-bedded (**LDTC-a**), often dewatered, argillaceous
228 sandstones (Table 1; Figure 7). Most beds include shell fragments and plant debris in the structured
229 intervals, and are capped by silty mudstones that regularly include bioturbations (Figure 7) such as
230 *Phycosiphon* sp., *Zoophycos* sp., *Skolithos* sp. and *Scolicia* sp. Episodes of thin-bedded bioclastic grits
231 (**sHDTC-b**) are recurrent (Figure 7). Very-thick sandstone beds punctually occur; they tend to have sole
232 marks, several grain-size breaks, and sometimes show decametric soft-sedimentation deformation
233 structures. A few lenticular, medium- to very thick-bedded sandstones (**HDTC**) are also observed. Their
234 incision rarely exceeds 50 centimeters and can comprise coarse bioclastic material (Figure 7).
235 Paleocurrent measurements from sole marks and traction structures generally indicate a north-eastward
236 direction at the base of this system (Figure 9). Finally, disorganized, contorted strata (**SL-a**) and coherent,
237 remobilized strata displaying sharp internal truncations in the same lithofacies (**SL-b**) recurrently interrupt
238 the strata (Table 1; Figure 6).

239 Interpretations

240 **Fa1g** is interpreted to record the emplacement and sustained development of a distal depositional lobe
241 system (Bailleul et al. 2007; Pr elat et al. 2009; McArthur et al. 2021) installed above and subsequently
242 pinned at its base within the rugose topography left by the underlying MTDs (e.g., Bull et al. 2009;
243 Armitage et al. 2009) (Figure 9; Figure 12a, b). These laterally extensive, tabular, sandstone-rich deposits
244 are interpreted as the deposits of dominantly low-density turbidity currents (**LDTC**) (Bouma 1962),
245 commonly deposited in off-axis to fringe positions (Pr elat et al. 2009). The punctual bioclastic grits as
246 well as the locally erosive, amalgamated and very thick occurrences, are however attributed to higher-
247 density turbidity currents (Lowe 1982) resulting from larger flow events. The onlapping and growth strata

248 geometries as well as the recurrent yet localized remobilization of strata both suggest an unstable
249 depositional environment, likely due to a continued rise of the eastern structure. The paleocurrent
250 directions to the NE indicate that the flow at the start of this depositional system was parallel to the rising
251 NE-SW oriented structure to the east and thus both point towards a different sourcing system to the
252 underlying MTDs (see Section 5.3.2) and an axial routing of the turbidity currents (Burgreen and Graham
253 2014). The abundance of land-derived material suggest that these turbidity currents likely initiated from
254 a shallow marine environment connected to a hinterland (Kuenen 1964), while the micropaleontology
255 analyses indicate that deposition of the turbidites occurred at mid bathyal depths or deeper (> 800 meters)
256 (Appendix 2).

257 6. SOURCE REGIONS

258 Despite lithofacies similarities, the sediment gravity flows observed at Waikaraka and Homewood are
259 lithologically different (Figure 13) and thus indicate different sediment sources. Here, we use the nature
260 of the reworked material in the conglomerates (**DF**, **CF**, **gHDTC**, **gsHDTC**), pebbly sandstones (**sHDTC-**
261 **a**), bioclastic grits (**sHDTC-b**) and coarser intervals of the sandstones (**HDTC**) as a tool for better
262 understanding and characterizing their provenance and source areas as well as deciphering local
263 paleotransport directions (e.g., Nemeč and Steel, 1984).

264 6.1. Extrabasinal sources

265 Clast observations

266 Across both Waikaraka and Homewood, the conglomeratic units (**DF**, **CF**, **gHDTC**, **gsHDTC**, **sHDTC**,
267 **HDTC**) essentially rework extraformational clasts from pre- and syn-subduction strata (>95%) (Figure
268 13). Pre-subduction material dominates (> 80%) and comprises lithoclasts from the Early Cretaceous up
269 to the Paleocene at Waikaraka and from the Late Cretaceous up to the Eocene at Homewood. The syn-
270 subduction material is divided into Miocene lithoclasts and bioclasts.

271

272

273

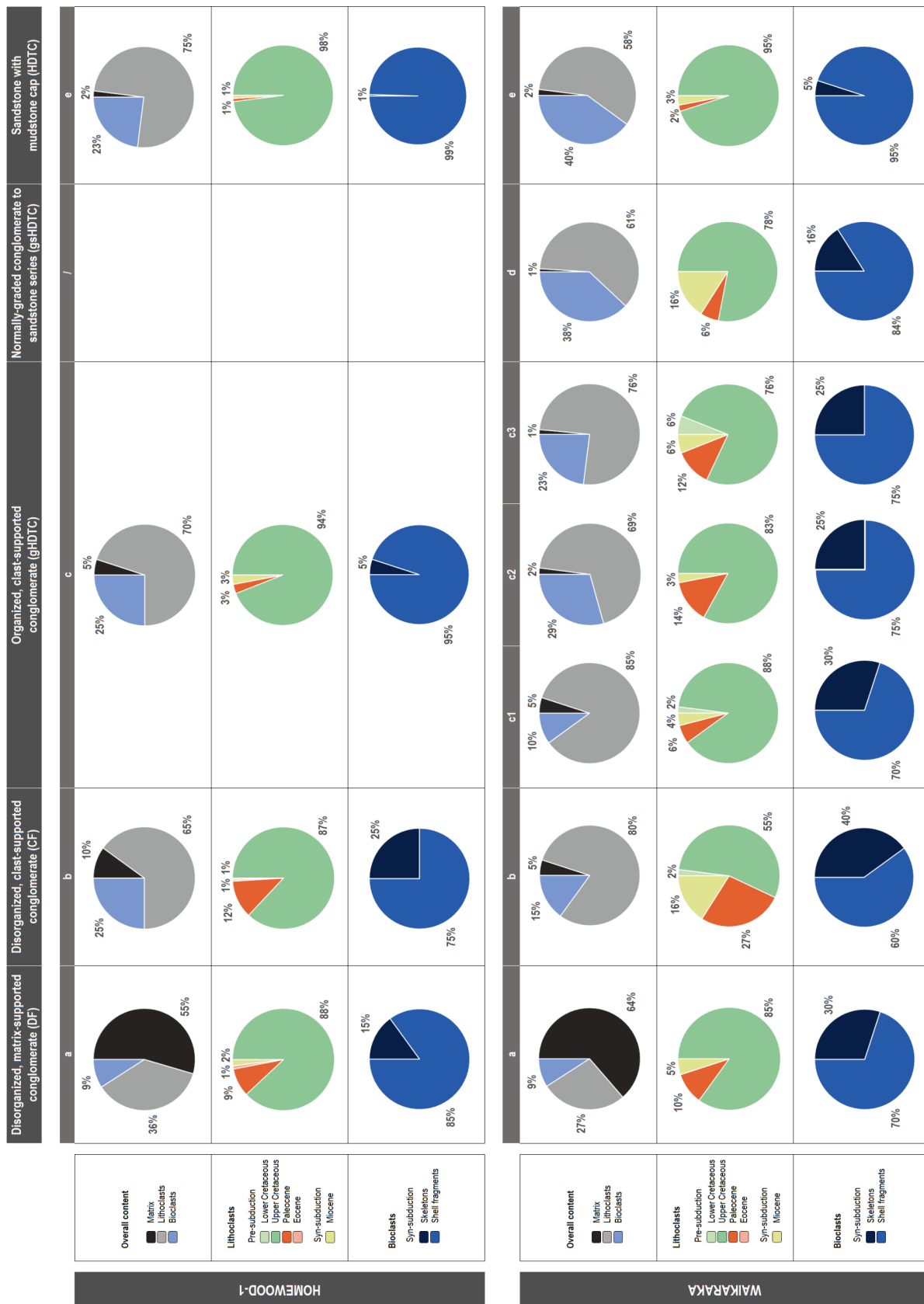


Figure 13: Quantitative analyses of the clast content recognized at Waikaraka and Homewood outcrops. The analyses looked at both the lithoclast and bioclast content. The lithoclasts were classified according to the tectonostratigraphic unit they

274 belonged to and the bioclasts were subdivided into two main types, *i.e.*, the preserved skeletons and the shell fragments. DF,
275 CF, gHDTC, gsHDTC and HDTC correspond to the lithofacies summarized in Table 1.

276 Intraformational clasts are virtually absent (<1%) in **CF**, but are more frequent in the conglomerates
277 deposits resulting from matrix strength (**DF**; *e.g.*, poorly lithified turbidites) and fluid turbulence (**gHDTC**,
278 **gsHDTC**, **sHDTC**, **HDTC**; *e.g.*, mud clasts).

279 Clast interpretations

280 The dominance of extraformational clasts implies that the flows responsible for these conglomeratic units
281 initiated outside of their final depositional environment and incorporated material belonging to structural
282 units and or paleogeographic domains different to that of the intrabasinal sediments (*i.e.*, extrabasinal)
283 (*e.g.*, Ogata et al. 2019). The abundance of pre-subduction material indicates that during the Middle
284 Miocene, the exposed extrabasinal paleogeography was dominated by Cretaceous to Eocene strata. In
285 compressional settings, uplift of thrust-related structures is known to develop bathymetric highs and or
286 steep slopes, which can in turn, expose rocks from the older substratum (*i.e.*, pre-subduction), ready to
287 be reworked and incorporated into the subsequent flows and MTDs (*e.g.*, Ogata et al. 2019).

288 **6.2. Mixed siliciclastic-carbonate shelves**

289 Bioclast observations

290 A substantial amount (~10 to 40%) of syn-subduction bioclasts is present in all the deposits, either as
291 shell fragments or as well- to partly-preserved skeletons (Figure 13). Molluscan species (*e.g.*, bivalves
292 and gastropods) dominates; yet, corals (solitary and colonial) and red algae are also locally present
293 (Appendix 3). At Homewood, the preserved forms of macrofossils reach between 15 and 25% of the
294 bioclasts contained in the debris flow deposits (**DF** and **CF**) whereas they are virtually absent in the
295 turbidites (**gHDTC** and **HDTC**), which are dominated by shell fragments. At Waikaraka however, the
296 preserved forms represent a considerably larger proportion of the bioclasts, mostly oscillating between
297 25 and 40%, even at the base of the turbidites (**gHDTC** and **HDTC**). Although the content in both the
298 coarse (macrofossils) and fine (shell fragments) skeletal fractions gradually increases toward the top of
299 the Waikaraka deposits, this upward evolution is particularly remarkable in the late-stage fill deposits
300 (**Fa2c-I**) (Figure 13, cf. c2, c3, d, e).

301 Overall, the taphonomic analyses shows that destructive processes dominate the taphonomic record of
302 the deposits, constructive taphonomic processes such as encrustation being virtually absent (Figure 14).
303 At Homewood, the deposits are characterized by bioclasts being: (1) moderately to highly fragmented,

304 (2) moderately abraded and, (3) poorly bioeroded. At Waikaraka, the finer bioclastic fractions compare
305 markedly in both debris flows (**DF**, **CF**) and high-density turbidite flows (**gHDTC**, **gsHDTC**). By contrast,
306 the coarser skeletal material in the former exhibit higher degrees of abrasion and bioerosion than in the
307 latter (Figure 14).

308 Bioclast interpretations

309 The nature and amount of reworked molluscan species indicate that the sediment gravity flows originated
310 from shallow marine middle- to inner-shelf neritic environments before being transported to deep-marine
311 outer to mid bathyal depths (Appendix 2; Appendix 3).

312 The contemporaneously developing mixed siliciclastic-carbonate environments (*i.e.*, compositional
313 systems (*sensu* Chiarella et al. (2017))) previously identified and described in the Coastal Ranges present
314 markedly similar faunal assemblages (Crundwell 1987; Chanier 1991; Bailleul et al. 2007; Bailleul et al.
315 2013; Bailleul et al. 2015) and are thus great candidates for providing both the preserved skeletons and
316 shell fragments. The depositional system envisaged for those sources is one of a regional, continent-
317 attached shelf domain, recurrently supplied with terrigenous sediments from the exposed Cretaceous
318 basement located to the west (Crundwell 1987; Chanier 1991; Bailleul et al. 2007; Bailleul et al. 2013;
319 Bailleul et al. 2015; Clausmann et al. 2021).

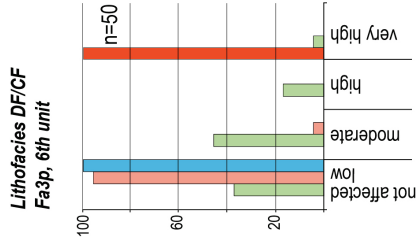
320 The substantial amount of gravels transported beyond the shelf and captured within the deposits (see
321 Sections 5 and 6.3) requires narrow shelf settings to easily connect to the slope beyond (Nemec and
322 Steel 1984), which typify the shelfal systems along active margins (Sømme et al. 2009). No lateral
323 continuity has yet been proven between the mixed siliciclastic-carbonate environments and therefore,
324 isolated, continent-detached shelfal systems likely also developed locally (*e.g.*, (Bailleul et al. 2013),
325 Fingerpost shelf) upon and around actively growing folds and faults (Caron et al. 2004) (see Section 6.3
326 and 7).

327 In fact, at Waikaraka, the fairly constant proportion of shelfal macrofossils found throughout the 90 meter-
328 thick succession points to paleophysiographies allowing for molluscan communities to thrive above
329 storm-wave base (Pomar 2001). At Homewood however, the macrofossils are mostly found in the
330 conglomerates (**DF**, **CF**) of the mass-wasting system (**Fa3p**) and their quantity tends to be generally
331 smaller to that of Waikaraka. This could indifferently indicate a smaller-sized source area, slightly different
332 paleoenvironmental conditions at the source as well as different transport processes and distances. The
333 paleotransport directions recorded in the mass-wasting system (**Fa3p**) of Homewood also inform that the

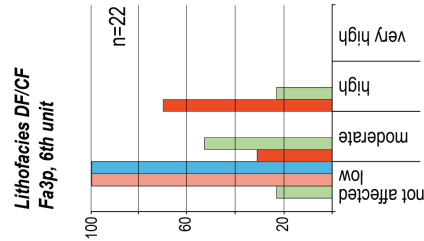
334 shelfal source was located to the east and thus was most-likely detached from the main regional shelf
 335 domain located to the west.

HOMEWOOD OUTCROP (Fa3p)

Bioclastic fraction (0.5 cm-3 cm)

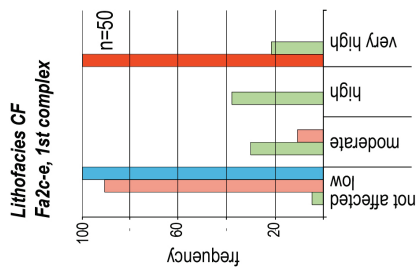


Bioclastic fraction (>3 cm)

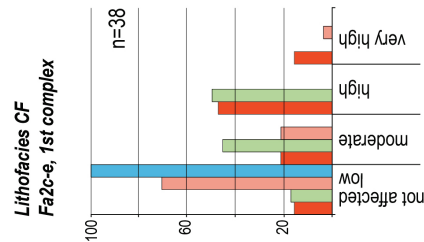
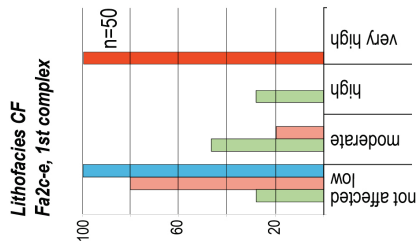
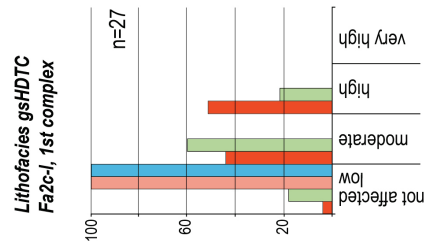
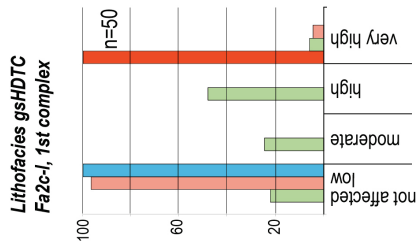
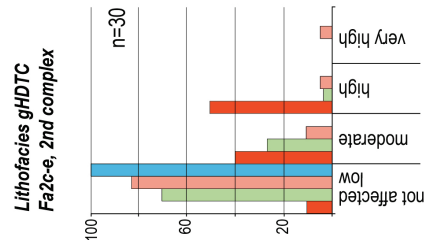
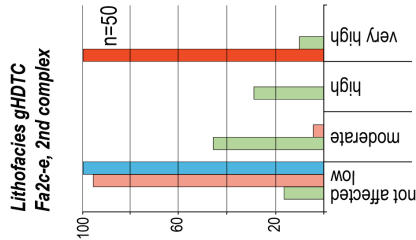
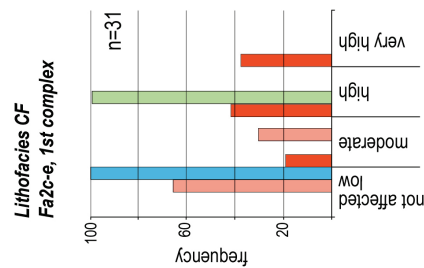


WAIKARAKA OUTCROP (Fa2c)

Bioclastic fraction (0.5 cm-3 cm)



Bioclastic fraction (>3 cm)



LEGEND

- Fragmentation
- Abrasion
- Bioerosion
- Encrustation

336

337 Figure 14: Taphonomic characterization of fossil remains described from one square meter area of outcrop at Waikaraka and
338 Homewood, using frequency histograms of the degree of alteration (e.g., low, moderate, high, very high) for each category of
339 skeleton damage (i.e., fragmentation, abrasion, bioerosion) and encrustation. Lithofacies CF, DF, gHDTC and gsHDTC are
340 described in details in Table 1 and facies associations Fa2c and Fa3p in Section 5.

341 Conversely, the profuse shell fragments found in the overlying, well-developed, north-eastward
342 prograding turbidite system (**Fa1g**) indicate a change of source region, likely belonging to the main
343 westward continent-attached shelf domain.

344 The regular detrital sand and pebble supply (see Sections 5 and 6.3) likely restricted the development of
345 carbonate factories dominated by suspension-feeding organisms (Caron et al. 2004), thereby promoting
346 molluscan communities (Bimol assemblage of Hayton et al. (1995)) often associated with relatively turbid
347 waters due to fine material in suspension. The occurrence in the deposits of occasional coralline red algal
348 branches and small-sized-rhodoliths indicate sources within the photic zone (Pomar 2001; Nebelsick et
349 al. 2005).

350 Differences in the degrees of taphonomic alteration observed in skeletal remains relate to their residence
351 time in the taphonomic active zone (TAZ; (Davies et al. 1989)). Typically, taphonomic alteration
352 consistently increases with increasing residence time in the TAZ where skeletons are affected by
353 destructive processes, such as disarticulation, fragmentation, abrasion and bioerosion. For example, low
354 net sedimentation rates and repeated exhumation of buried skeletal remains by waves and currents, and
355 by seafloor burrowers favor mechanical destruction and bioinfestation. By contrast, rapid and definite
356 burial prevents such alterations (e.g., Best and Kidwell 2000; Best et al. 2007). Fragmentation of skeletal
357 remains over three centimeters in size is moderate to high, both at Waikaraka and Homewood in debris
358 flows and turbidite flows (Figure 14). Although it cannot be ruled out that skeletons were partly fragmented
359 in their source area through hydraulic and biological processes (Zuschin et al. 2003), it should be
360 expected that fragmentation also occurred, as a result of mechanical crushing, once they were integrated
361 in their host clast-supported gravity-driven flows. Abrasion of the coarser skeletal fraction, translating in
362 the loss of ornamentation and rounding of bioclast edges, is: (1) higher in debris flows than in turbidite
363 flows, and (2) overall higher than fragmentation in debris flows compared to turbidite flows. Abrasion is
364 caused by physical erosion, through friction and at a lesser degree through waterborne collision, and
365 reworking leading to mobilization and redeposition of bioclasts. Moderate to high degrees of abrasion,
366 along with the abundance of small-sized skeletal debris, are indicative of frequently destabilized
367 substrates by wave action, storms and, potentially here, gravity-driven flows (e.g., Kowalewski 1996).
368 Further, such unstable substrates may have inhibited colonization of dead shells by encrusters, and their
369 infestation by borers, thereby explaining the low degrees of taphonomic alteration for these processes

370 (Figure 14). However, possible causes for the observed differences in biogenic alteration (slightly higher
371 in Waikaraka debris flow deposits for example) are multifarious, including high-sedimentation rates
372 leading to rapid burial, unsuitable shell substrates and/or ecological conditions for infester communities
373 (Caron et al. 2019).

374 Results of the taphonomic evaluation suggest that the alteration state of skeletal remains in the studied
375 deposits not only reflects depositional and ecological conditions at the site of carbonate production on
376 the shelf, but has also potential to record conditions of transport during their remobilization in gravity-
377 driven systems. In particular, the taphonomy of skeletal contents, as a complementary tool to other
378 analytical approaches, may provide clues on the laminar (*i.e.*, debris flows) versus turbulent (*i.e.*,
379 turbidites) character of flow, and whether deposition by traction sedimentation prevailed over suspension
380 sedimentation (Lowe 1982).

381 **6.3. Distinct substratum and continental connection**

382 Lithoclast observations

383 At Homewood, the extra- and intraformational lithoclasts tend to oscillate between sub-angular to sub-
384 rounded shapes and are in average of granule- to pebble-grades in the confined channel fills (**Fa2c**) and
385 granule- to cobble-grades in the mass-wasting system (**Fa3p**). Conversely, at Waikaraka, sub-rounded
386 clasts dominate and sub-angular shapes only preferentially occur in the early-stages confined channel
387 fills (**Fa2c-e**). At Waikaraka, the clast grades also largely depend on the style of channel fills, and are
388 preferentially of pebble- to outsized-grades in the early-stages fills (**Fa2c-e**) and of pebble-grades in the
389 late-stage fills (**Fa2c-l**) (see Section 5).

390 Pre-subduction extraformational clasts largely dominate representing 95 to 99% of the overall lithoclast
391 content at Homewood and 84% to 97% at Waikaraka (Figure 13). They mostly rework (>80%) Upper
392 Cretaceous strata (*e.g.*, mainly siliceous and calcareous mudstones, rare lavas) and frequently include
393 (~4 to 10%) Paleocene clasts (*e.g.*, limestones). Conversely, Lower Cretaceous clasts (*e.g.*, Torlesse
394 greywackes) are only observed, sporadically (~2%), at Waikaraka, whereas Eocene clasts (*e.g.*, smectitic
395 mudstones) are virtually absent and only rarely found (~1%) at Homewood (Figure 13).

396 At Waikaraka, the early-stage fills of the first channel complex (**Fa2c-e**) deviate from this model,
397 remobilizing sub-angular to sub-rounded, pebbles to outsized clasts (deci- to decametric) from the entire
398 Lower Cretaceous to Paleocene pre-subduction series at their base. Interestingly, sub-rounded to
399 rounded, granule- to pebble-grade Lower Cretaceous clasts are also occasionally observed throughout

400 the confined slope channel system (**Fa2c**) (Figure 13). They are preferentially restricted to particular
401 conglomerate intervals (**CF**, **gHDTc**) of the confined channel fills, generally located at the top of the early-
402 stage fills (**Fa2c-e**).

403 The syn-subduction extraformational lithoclasts most commonly include pebble- to boulder-grade clasts
404 of lithified shell beds. They are frequent in the confined slope channel system of Waikaraka. There, eight
405 distinct varieties of shell bed clasts are observed (Appendix 5), which are often of cobble- to boulder-
406 grades in the early-stage fills (**Fa2c-e**). Thick-shelled Molluscan remains are frequent in the shell bed
407 clasts (Appendix 5, A-F, H), along with well-rounded lithic pebbles commonly bearing the trace fossil
408 *Gastrochaenolites* (Appendix 5, B-D). The overall faunal assemblages, including *Struthiolaria* sp.,
409 *Polinices* sp., Oysters, Turritellids, Serpulids and zooxanthellate corals (Appendix 5), and moderate
410 degrees of pre-burial taphonomic alteration collectively point to rapid deposition in high-energy proximal
411 inner-shelf to middle-shelf environments. Conversely, shell bed clasts are only punctually found (<0.5%)
412 in the confined slope channel (**Fa2c**) and mass-wasting system (**Fa3p**) of Homewood. They however
413 only comprise pebble- to cobble-grade skeletal and bioclastic fine-grained sandstones dominated by
414 turritellid gastropods (cf. facies Fa6c from [Bailleul et al. \(2007\)](#)).

415 The remaining syn-subduction extraformational lithoclasts comprise pebble-grade wood fragments and
416 disseminated plant debris. A few cobble- to boulder-grade clasts of fossil tubular concretions ([Malie et al.](#)
417 [2017](#)) were also observed in the early-stage fills (**Fa2c-e**) at Waikaraka.

418 Finally, the syn-subduction intraformational lithoclasts are characterized by outsized, contorted
419 sandstones and pebble- to boulder-grade mudstone clasts. The outsized, contorted sandstones are
420 generally restricted to **DF**, particularly at Homewood; whereas the mudstone clasts, although present in
421 some of the conglomerates, are dominantly found in the coarse-grained basal intervals of the sandstones
422 at both Waikaraka and Homewood.

423 Lithoclast interpretations

424 As previously observed elsewhere in the Coastal Ranges ([Neef 1992](#); [Bailleul et al. 2013](#); [Claussmann](#)
425 [et al. 2021](#)), the nature of the extraformational lithoclasts informs that the mixed siliciclastic-carbonate
426 shelfal domain sourcing the Waikaraka deposits developed above a pre- (e.g., Cretaceous to Paleogene
427 series) and syn-subduction (e.g., Miocene sediments) substratum, whereas the paucity of
428 extraformational syn-subduction clasts (e.g., a couple of shell bed clasts) at Homewood suggests a pre-
429 subduction-dominated substratum.

430 At Waikaraka, the variety encountered in both the proportion and nature of pre-subduction material
431 informs that the basal deposits of the confined slope channel system (**Fa2c**) remobilized the entire pre-
432 subduction series down to the Lower Cretaceous, whereas the remainder mostly captured remobilized
433 Upper Cretaceous strata (Figure 13). The absence of Lower Cretaceous greywackes and predominance
434 of Upper Cretaceous strata at Homewood both suggest a slightly different substratum and or exhumed
435 paleobathymetric high sourcing the deposits. The implications for both these localities are discussed in
436 Section 7.

437 At Waikaraka, the syn-subduction material, mostly comprising shell bed clasts, likely results from
438 destabilization of previously deposited and lithified strata (*i.e.*, substratum) rather than contemporaneous
439 sediments. Their diversity captures different mixed siliciclastic-carbonate shelfal environments, which
440 either (1) belonged to the same shelfal system, thus suggesting different phases of shelf development or
441 local variations in depositional settings (Bailleul et al. 2015), and or (2) resulted from distinct shelfal
442 systems that formed in the area since the onset of subduction (Bailleul et al. 2007; Bailleul et al. 2013;
443 Caron et al. 2021). Conversely, at Homewood, the occasional sandy shell beds could be
444 contemporaneous with the development of the shelfal environment. Early lithification of the deposits at
445 or near the seafloor may have been triggered by increased saturation of seawaters with respect to
446 CaCO₃, following partial dissolution of aragonitic turritellid shells, though such process has been
447 documented to predominantly result from burial diagenesis (*i.e.*, mesodiagenesis) (Nicolaidis 1995;
448 Haywick 2004; Caron and Nelson 2009).

449 The abundant pebble-grade to outsized sub-angular clasts found in the confined slope channel (**Fa2c**)
450 and mass-wasting system (**Fa3p**) of Homewood, and more sporadically in the early-stage confined
451 channel fills (**Fa2c-e**) of Waikaraka, both indicate that despite clast interactions, not only the sediment
452 sources were nearby but also that the material was “freshly” eroded to allow preservation of such features
453 (Postma et al. 1988).

454 Notwithstanding partial remobilization of the shelf and slope substratum through submarine
455 destabilization (*e.g.*, mass-wasting failure) and or erosional processes (*e.g.*, channel cutting through the
456 slope substratum) to provide for most of the lithoclast content (*e.g.*, Sømme et al. 2009; Janocko et al.
457 2013; Paumard et al. 2020), dismantlement and erosion of exhumed paleobathymetric highs probably
458 contributed as well (*e.g.*, Ogata et al. 2019). In fact, the presence of wood fragments and plant debris in
459 both settings informs that the source areas were connected to a terrestrial ecosystem (Kuenen 1964),
460 which indicate the nearby presence and connection with exhumed land. Also, the occurrence of sub-
461 rounded lithoclasts can suggest a fluvial and or shallow-marine origin with extraformational clasts that

462 resided for at least some time in a littoral zone, before being transferred to bathyal depths (Di Celma et
463 al. 2010). Where grain collision is inferred as a dominant grain-support mechanism (e.g., in most of the
464 **Fa2c** and **Fa3p** deposits), it can also be expected to have favored the quick abrasion and rounding of the
465 harder, coarse-grained clasts.

466 The large volume of wood fragments and plant debris recorded throughout the confined slope channel
467 deposits of Waikaraka (**Fa2c**) and the well-developed turbidite system of Homewood (**Fa1g**) both support
468 source connections to mature hinterlands (e.g., Stanley 1986; McArthur et al. 2016b). The discrete
469 occurrences observed in the mass-wasting system of Homewood (**Fa3p**) rather points toward a small
470 and or isolated source system, such as an island developed above tectonically-controlled topography
471 (McArthur et al. 2016a), hence reinforcing the hypothesis of a source system detached from the main
472 shelf domain.

473 Finally, the nature of intraformational clasts, such as mud clasts, informs that these gravity flows were
474 erosive and ploughed through mud-prone substrates as they moved downslope or at hydraulic jumps
475 (Posamentier and Martinsen 2011). At Homewood, the contorted sandstones also indicate of the varying
476 nature of the ploughed substrate, locally supported by the underlying sandier deposits (**Fa2c**).

477 7. DISCUSSION

478 7.1. Gravity-driven systems from forelimb, continent-attached shelfal 479 source

480 The depositional system proposed for Waikaraka is one of a confined slope channel system (CSCS)
481 (*sensu* Campion et al. 2000) distributing land-derived, shallow- and deep-marine bedrock-derived
482 sediments to the trench-slope basin.

483 As interpreted (see Sections 5 and 6), this system was likely sourced from the regional, narrow and
484 continent-attached mixed siliciclastic-carbonate shelfal domain, which was contemporaneously
485 developing to the west during the Middle Miocene, atop seaward-verging thrust-cored anticlines (Figure
486 15A; Figure 16) (Crundwell 1987; Chanier 1991; Bailleul et al. 2007; Bailleul et al. 2013; Bailleul et al.
487 2015; Clausmann et al. 2021).

488 Precise geometry of the slope upon which the CSCS developed is here a matter of speculation; yet, the
489 forelimb settings rather suggest a steep, probably irregular, tectonically-controlled gradient close to the

490 shelf edge and across the slope (Figure 15A; Figure 16) (e.g., Naranjo-Vesga et al. 2020). The large
491 volume of terrigenous material as well as the nature of the reworked clasts (e.g., well-preserved bioclasts,
492 shell fragments, pre-subduction strata) recorded throughout the CSCS deposits (**Fa2c**) both indicate
493 persistent connection with a shelfal domain, recurrently supplied by the mature, vegetated hinterland to
494 the west, exhumed since the onset of the Hikurangi subduction and mostly exposing Cretaceous strata
495 (Figure 16) (Chanier and Ferrière 1991).

496 Along active margins, CSCS commonly represents long-lived transport routes for sediments across the
497 shelf and slope that are typically insusceptible to sea-level changes and directly connect the hinterland
498 (Type I, shelf-incising and river-associated system of Harris and Whiteway (2011)) and or shoreface
499 depositional systems (Type II, shelf-incising system of Harris and Whiteway (2011)) to the deep sea
500 (Sømme et al. 2009). Here however, although regional tectonics (i.e., margin uplift) largely governed the
501 initial emplacement of this mixed siliciclastic-carbonate shelfal domain installed atop thrust-cored
502 anticlines (Bailleul et al. 2013), eustatic processes appear to then chiefly control their development and
503 stratigraphic architecture (Bailleul et al. 2015). Notwithstanding the reworking of Upper Cretaceous
504 conglomerates characteristically rich in already-reworked Lower Cretaceous Torlesse greywackes
505 (Chanier and Ferrière 1991) as a potential source for providing gravels readily available to be transferred
506 across the narrow shelf, the punctual input of well-rounded pebbles of Torlesse greywackes observed at
507 the top of the early-stages fills of the CSCS (see Section 6) could also be attributed to an intermittent
508 connection of the shelf with river(s). The successive T-R sequences that characterize these shelfal
509 environments (Bailleul et al. 2015) could explain such periodic communication with the exhumed
510 Cretaceous hinterland and would in turn suggest some eustatic control here, eventually captured in the
511 CSCS infill.

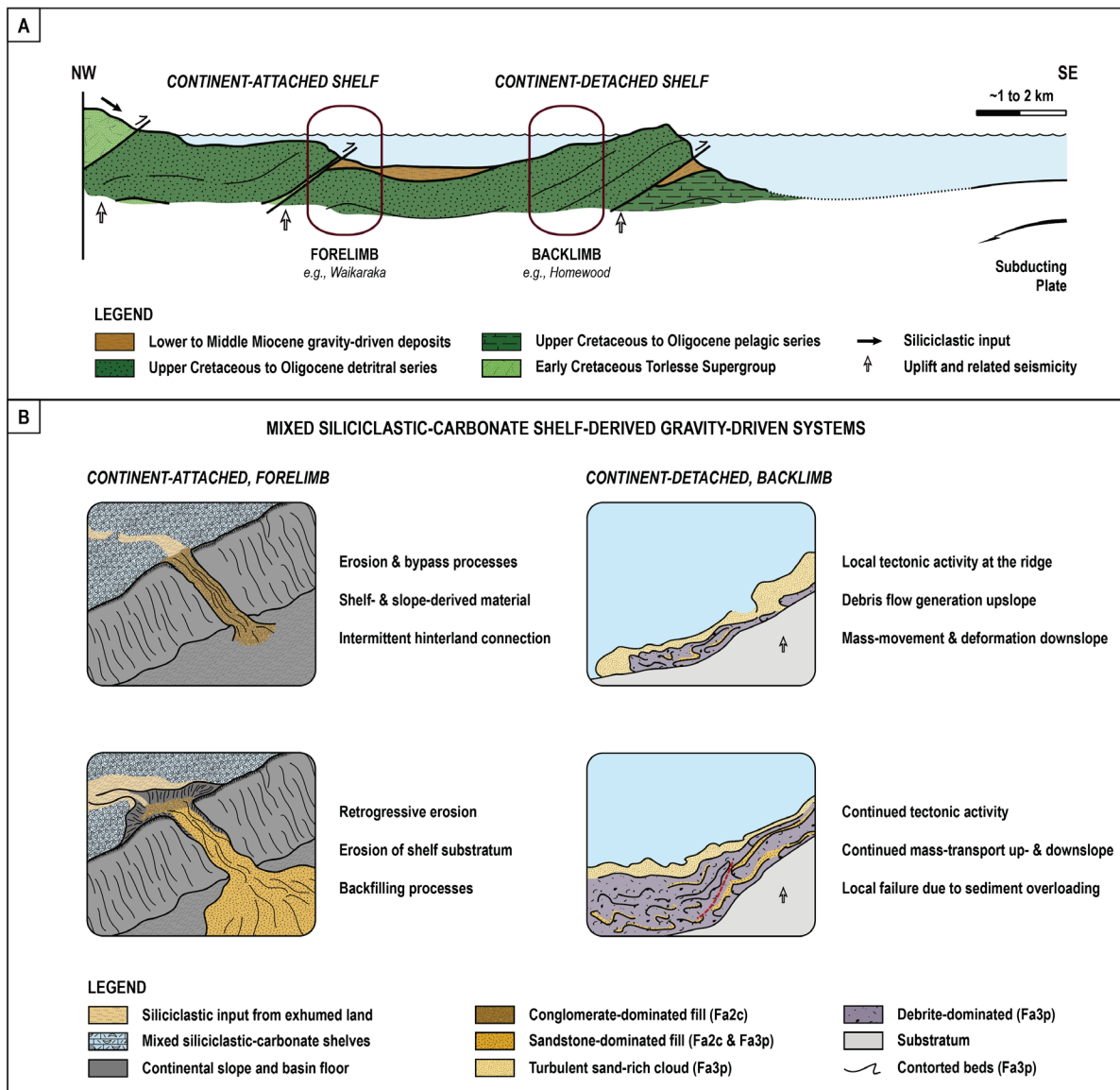
512 The broad fining- and thinning-upwards trends observed in the CSCS (**Fa2c**) are common in many other
513 slope channel systems, indicating channel bypass and backfilling processes (e.g., Mutti 1985; Deptuck
514 et al. 2003; Beaubouef 2004; Sylvester et al. 2011; Kneller et al. 2020). The repeating series of
515 architecture and facies associations inform that this CSCS system likely behaved in a cyclic manner
516 through time that may well have been in phase with the sea-level changes, yet likely also resulted from
517 external forcing factors, such as pulses of thrust fault activity (e.g., uplift and or related seismicity) at the
518 edge of the shelf, known to periodically renew erosion and active sediment transport into CSCS along
519 active margins (e.g., Mountjoy et al. 2009; Mountjoy et al. 2018; Watson et al. 2020).

520 Interestingly, the reworked material contradicts the traditional unroofing model, incorporating clasts from
521 the entire Lower Cretaceous to Paleogene pre-subduction strata only within its lowermost record and

522 being otherwise dominated by Upper Cretaceous strata (see Section 6.3). Although direct erosion of the
523 exhumed Cretaceous basement likely acted as an additional source, we attribute the former clast content
524 to chiefly result from CSCS forming processes, such as initial slope failure (*e.g.*, Janocko et al. 2013) and
525 or slope erosion (*e.g.*, Sømme et al. 2009). These processes exposed the underlying submarine bedrock
526 of the thrust-cored anticline (*i.e.*, forelimb setting) upon which the shelfal domain developed, making it
527 ready to be included into the subsequent gravity flows (Figure 15A; Figure 16) (*e.g.*, Ogata et al. 2019).
528 For the latter contrasting clast content, we here infer the autogenic evolution of this shelf-incising CSCS
529 (Type I, shelf-incising and river-associated system of Harris and Whiteway (2011)) during late low-stand
530 conditions, whereby retrogressive erosion at the head of the system (*sensu* Daly 1936) would have
531 favored the failure, storage and reworking of the marginal siliciclastic-carbonate sediments,
532 contemporaneously (*e.g.*, living or freshly deceased macrofauna, terrigenous debris) and previously
533 deposited (*e.g.*, shell bed lithoclasts), along with their underlying substratum, here dominated by Upper
534 Cretaceous strata (Figure 15B; Figure 16). The upward increase of bioclastic material recorded in the
535 channel fill may also partially illustrate such progressive upslope variation in source region, whereas the
536 dominance of skeletal grains (over non-skeletal, *e.g.*, ooids, pellets) typically occur during such sea-level
537 low-stands (Reijmer et al. 2015).

538 Here, regional tectonics appear to have chiefly governed both the birth (emplacement) and death
539 (drowning) of (1) the continent-attached mixed siliciclastic-carbonate shelfal domain installed atop thrust-
540 cored anticlines (Bailleul et al. 2013; Bailleul et al. 2015) and of (2) its related shelf-derived, deep-marine
541 systems. Whilst the shelfal domain development then became largely eustatically-controlled (Bailleul et
542 al. 2015), the combination of steep slope gradient, high sediment supply and narrow shelf width resulted
543 in the development of a CSCS (**Fa2c**) that efficiently transferred the shelf-derived sediments to the deep-
544 marine environments (*e.g.*, Naranjo-Vesga et al. 2020). Although punctually capturing the sea-level
545 changes, the overall evolution and geometry infill of the CSCS seemed essentially controlled by local
546 tectonic activity at the mixed siliciclastic-carbonate shelf edge, which, in turn, determined its stratigraphic
547 infill (sediment distribution and architecture) (Figure 15; Figure 16).

548 The abrupt changes in sedimentation styles (*i.e.*, sand-dominated) and paleodepth of deposition (*i.e.*,
549 lower bathyal) recorded in the uppermost CSCS deposits (see Section 5; Appendix 2) could be attributed
550 to the major change in tectonic regime (*i.e.*, generalized subsidence) that was recorded along the
551 Hikurangi margin at the time, which in turn resulted in rapid bathymetric deepening of the area (*i.e.*, high-
552 stand conditions) (Chanier et al. 1999; D5 discontinuity and U5 unit of Bailleul et al. 2013) (Figure 10a,
553 e). Direct coarse-grained sediment supply stopped; yet, some high-stand activity was maintained along
554 the CSCS (*e.g.*, Mountjoy et al. 2009).



555

556 Figure 15: (A): Schematic NW-SE cross-section through the subduction-related, compressive Whareama intra-slope basin

557 (*i.e.*, trench-slope basin) during the Middle Miocene. The depositional settings for the forelimb and backlimb settings of actively

558 growing thrust-cored structures are highlighted. (B): Schematic representation of the inferred shelf-derived gravity-driven

559 systems and processes related to the forelimb and backlimb settings.

560

561 7.2. Gravity-driven systems from backlimb, continent-detached shelfal 562 source

563 The source and depositional system envisaged at Homewood contrast with the coeval continent-attached

564 shelfal environment and slope channel delivery system of Waikaraka. As previously described (see

565 Sections 5 and 6), the nature of the deposits (**Fa2c** and **Fa3p**) indicates a shelfal environment that

566 developed upon and around a structurally-controlled ridge located to the east of Homewood, which was
567 detached from the regional shelf and hinterland (Figure 15A).

568 Isolated platforms may develop above fault-growing folds and favor biogenic carbonate production as
569 they lack continent-derived siliciclastic inputs (Caron et al. 2004; Ćosović et al. 2018; Caron et al. 2021).
570 At Homewood however, analyses of the reworked material (see Section 6) inform that mixed siliciclastic-
571 carbonate sediments dominated, thereby suggesting both emergence and subsequent erosion of the
572 uplifted ridge (Figure 15A; Figure 16). Previous work highlighted that islands developed above
573 tectonically-controlled topography can supply terrigenous material (e.g., McArthur et al. 2016a) and result
574 in mixed siliciclastic-carbonate sedimentation (e.g., Chiarella et al. 2012; Chiarella and Longhitano 2012;
575 Ćosović et al. 2018).

576 More particularly, the nature of the deposits (i.e., **Fa2c** and **Fa3p** slope deposits) and their clasts (see
577 Part 5) suggests that the ridge was nearby and exhumed for some time to allow (1) the dismantlement of
578 the pre-subduction series down to the Upper Cretaceous strata, (2) their subsequent reworking in the
579 littoral domain and (3) the formation of sandy shell beds. Notwithstanding submarine erosion, the
580 abundance of sub-rounded clasts of Upper Cretaceous strata implies that these strata likely outcropped
581 at the time (e.g., Di Celma et al. 2010), thereby inferring initial rise and emergence of the ridge prior to
582 the Middle Miocene, Upper Clifdenian to Lower Lillburnian (Langhian, 15.9 – 14 Ma, Appendix 2).

583 When combined, the pre-subduction Paleogene cover, about 600 meters thick in the study area (Van den
584 Heuvel 1960; Chanier 1991; Chanier and Ferrière 1991), and the general exhumation rate of up to 0.3
585 kilometer/million year calculated by Jiao et al. (2014) along the eastern North Island of New Zealand
586 indicate that the ridge emergence (i.e., island setting) and subsequent erosion started at least ca. 2 Ma
587 prior to that time. This ridge likely corresponded to the seaward margin of the Whareama Basin,
588 structurally controlled by the underlying seaward-verging Flat Point-Whakataki Fault complex and located
589 to the east of Homewood (i.e., Glenburn Nappe) (e.g., Chanier 1991) (Figure 3; Figure 15A). The set of
590 morphometric relationship equations calculated by Moscardelli and Wood (2015) predicts a length of
591 about two kilometers for the MTDs (Appendix 6), which may or may not include the evacuation area
592 (*sensu* Moscardelli and Wood 2015) yet provides a broad insight as to the potential distance to the ridge
593 at the time (less than two kilometers).

594 The envisaged shelfal system was narrow and located on the western side of the island, upon the
595 backlimb of the underlying asymmetrical thrust (Figure 15A; Figure 16). Similarly, the mixed siliciclastic-
596 carbonate Fingerpost shelf developed further north during the Early to Middle Miocene; it was also
597 positioned to the west of an emergent landmass (i.e., the Cape Turnagain Structural High), itself

598 controlled by an underlying seaward-verging thrust (*i.e.*, backlimb setting) (Bailleul et al. 2013, their figure
599 12).

600 At Homewood, the shelf was likely cut by a network of small slope channels connecting the island through
601 the shelf to the Whareama Basin (**Fa2c**) (Figure 16). The transition from slope channel (**Fa2c**) to shelf-
602 derived mass-wasting (**Fa3p**) systems is attributed to a period of tectonic activity (*e.g.*, uplift) at the basin-
603 bounding structure, resulting in growth strata deposition on the eastward-margin of the basin (see Section
604 5.3) (*e.g.*, Moore and Karig 1976; Stevens and Moore 1985) and typically lasting 1 to 2 Ma in these
605 settings (Nicol et al. 2002; Bailleul et al. 2013).

606 Structure growth both (1) promoted the repeated destabilization of the isolated shelf and generation of
607 the **Fa3p** mass-wasting flows as well as (2) controlled the axial routing and syn-kinematic development
608 of the overlying **Fa1g** turbidite lobe system sourced from the mature hinterland to the west (Figure 16).
609 The associated MTDs (**Fa3p**), only ~15 meters thick, recorded a difference of ~40° in bed dips (*i.e.*,
610 growth strata), thereby indicating complex interplay between the growing structure and sediment
611 deposition. The slope gradient acted as a dynamic control triggering (1) mass-wasting processes
612 upslope, as well as, down the slope, (2a) rapid flow transformation and efficient dilution (*e.g.*, Fisher 1983;
613 Strachan 2008) resulting into fast-moving multiphase debris flows (*e.g.*, Postma 1986; Mohrig et al. 1998;
614 Sohn et al. 2002) and (2b) coeval remobilization (soft-sediment translation and deformation) of the
615 previously deposited, unlithified mass-flow deposits (*e.g.*, Maltman 1994) (Figure 15B; Figure 16).

616 The MTDs' succession and the variations in slope gradient recorded throughout can therefore be used
617 as a proxy for the tectonic activity of an individual structure. Here, they inform that at least six main
618 tectonic pulses with uplift affected the structurally-controlled shelf-margin(s), each respectively
619 generating one to several coalescing debris flows. Towards the top of the MTDs' succession, the high
620 amount of sheared debris flow deposits, contorted strata, and the increasing bed dip variations (Figure
621 9) suggest both steep slope gradients (*e.g.*, Postma 1986) and an acceleration of the ridge activity. Steep,
622 rising slopes, recurrently loaded with sediments, are unstable (Hein and Gorsline 1981) and thus likely
623 provided the right settings for locally triggering the failure of the entire MTDs' series as previously
624 observed at Homewood (Figure 9; Figure 15).

625

7.3. Generic implications

626

7.3.1. For shelf-derived gravity-driven systems sourced from thrust forelimb

627

and backlimb

628 In this study, the high variability encountered within one intra-slope basin in terms of sediment delivery
629 systems as well as transport and depositional processes highlights the role and importance of varying
630 structural settings of the sediment source for controlling the subsequent deep-marine depositional
631 settings. Despite similar source environments (*i.e.*, mixed siliciclastic-carbonate shelves) developed atop
632 the basin-bounding structures (*i.e.*, thrust) along with comparable sediment delivery systems (*i.e.*, slope
633 channel and remobilization) and nature of flow processes (*i.e.*, gravity-driven), the resulting shelf-derived
634 deposits downslope are characterized by contrasting lithofacies assemblages, sizes and geometries.

635 Notwithstanding the size and extent of the catchment area or continental connection (attached or
636 detached) as critical controls (Sømme et al. 2009), we here suggest that along active margins, the
637 geometry and activity of the underlying structures (thrust forelimb and backlimb settings) at the shelf
638 edges also greatly influence the resulting sediment distribution and architecture.

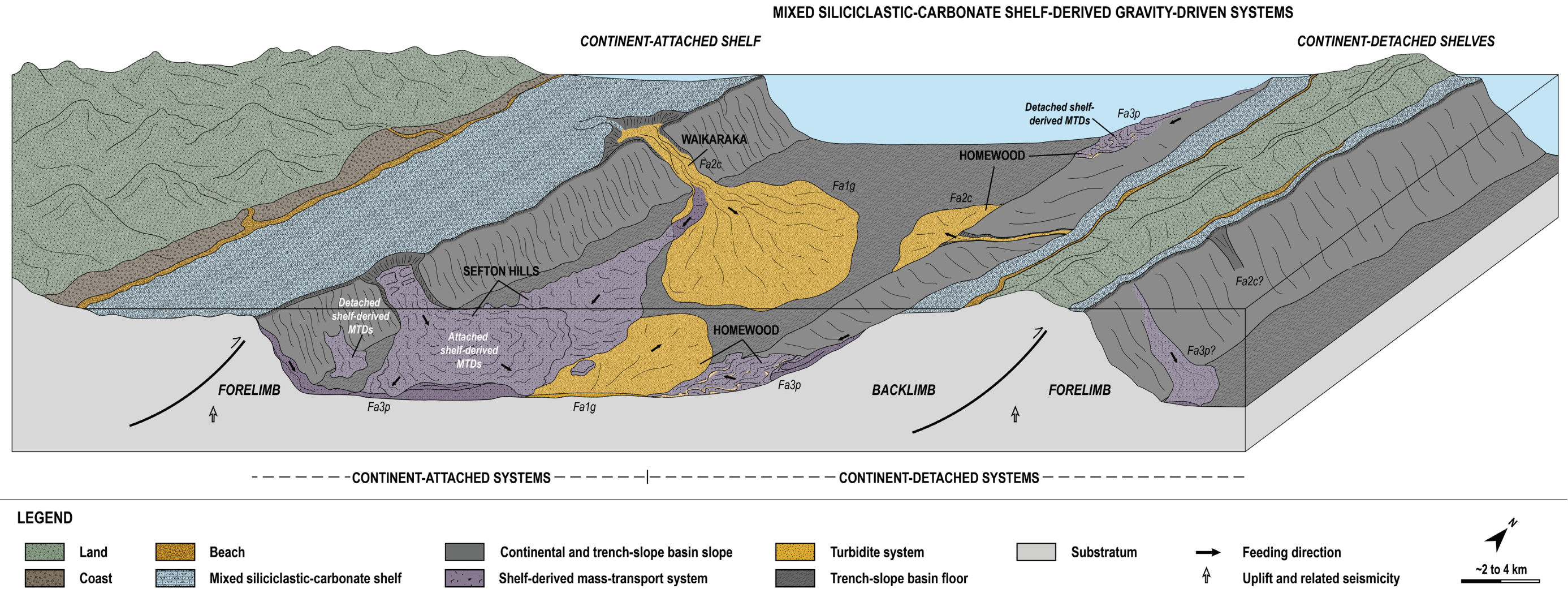
639 The shelf-edges on the forelimbs of underlying thrusts are abrupt, generally associated with steep,
640 irregular and tectonically-controlled slope profiles (Naranjo-Vesga et al. 2020). Erosion and transport
641 capacities proportionally increase with slope gradient and thus favor significant sediment bypass to the
642 deeper settings (Sømme et al. 2009; Crisóstomo-Figueroa et al. 2020). Here, the tectonic activity at the
643 shelf edges generally favored the development of systems that dynamically interact and rejuvenate the
644 slopes (*e.g.*, Waikaraka retrogressive system or Sefton Hills repeated collapse events (Clausmann et
645 al. 2021)) and the resulting gravity-driven systems are generally hundreds of meters thick and several
646 kilometers wide, preferentially transporting sediments through sediment gravity flows.

647 Conversely, thrust backlimb commonly provide longer and gentler slope gradients upon which a more
648 gradual shelf edge transition may be expected and may facilitate the development of less efficient
649 depositional systems (Sømme et al. 2009). In this study however, these settings were connected to
650 limited catchment areas (*e.g.*, isolated island), thereby preventing the development of particularly large
651 systems that typically occur on gentler slopes. Therefore, despite similar delivery systems to those
652 associated to forelimb-related slopes (*e.g.*, slope channels and destabilization), their sizes were largely
653 smaller (tens of meters thick).

654

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658

659

Figure 16: Contrasting depositional systems for mixed siliciclastic-carbonate shelf-derived gravity-driven flows along tectonically active margins, using the southern portion of the Hikurangi subduction margin as an example (Whareama trench-slope basin). Fa1g – refers to depositional lobe system; Fa2c – to confined slope channel system; Fa3p – to shelf-derived mass-wasting systems. Depositional settings from the Sefton Hills localities were adapted from Clausmann et al. (2021).

660 Destabilization and collapses of the mixed siliciclastic-carbonate shelfal sources occurred on both sides
661 of the basin-bounding structures and thus indicate that any of these shelves, either attached to the main
662 continental domain or detached from it (e.g., isolated island) can be subjected to shelf-derived sediment
663 mass movement and mobilization (*sensu Iverson 1997*).

664 When sourced from the backlimb of an actively growing structure (i.e., thrust) however, the mass-wasting
665 deposits exhibit more complex internal architectures, clearly recording the slope gradient variations and
666 resulting in intricate interactions between mass movements and mass flows, such as the ones observed
667 at Homewood. The changes in slope gradient will dynamically control both (1) the generation of mass
668 movements and flows upslope, and (2) the coeval remobilization (e.g., sliding) and soft-deformation (e.g.,
669 slumping) of the previously deposited, poorly-lithified sediments (e.g., debrites) downslope. Conversely,
670 although mass movement (e.g., slide, slump) can undoubtedly occur as well, the MTDs sourced from the
671 forelimb of an actively growing structure (i.e., thrust) seem primarily characterized by a series of
672 coalescing mass-flow deposits (e.g., debris flows) (e.g., Festa et al. 2010 and references therein). Mass
673 movements either seem preferentially recorded close to the source region or at the base of the MTDs'
674 sequence, where they often result from soft-sediment deformations (e.g., slide, slump) of the unlithified
675 sediments that were previously deposited at the front of the rising structure (e.g., Festa et al. 2010; Lackey
676 et al. 2020).

677 Overall, the shelf morphology (i.e., narrow), extent (e.g., regional or isolated), continental connection
678 (e.g., attached or detached), sediment type (i.e., mixed siliciclastic-carbonate sedimentation) as well as
679 the respective geotectonic settings (e.g., upon and around the forelimb or backlimb of the underlying
680 thrusts and fault-growing folds) all chiefly controlled the architecture and subsequent shelf-derived
681 deposits. However, along tectonically-active margins, when sourced from mixed siliciclastic-carbonate
682 systems, the deep-marine, shelf-derived deposits seem characterized by tens to a few hundred of meters
683 of thickness and several kilometers of lateral extent. The durations of the associated depositional systems
684 are from 1 to 2 Ma, being controlled by the structural styles and local tectonic activity at the source rather
685 than regional-scale variations.

686 **7.3.2. For mass-transport deposit nomenclature**

687 Moscardelli and Wood (2015) have proposed differentiating mass-wasting products into *attached* and
688 *detached* systems, respectively sitting in a proximal and a distal position from any shelf break, and either
689 sourced from this regional area (e.g., shelf) or local structures. *Attached* MTDs commonly represent the
690 largest occurrences, being regionally extensive, and are essentially triggered by sea-level variations and

691 or high sedimentation rates (Moscardelli and Wood 2008; Posamentier and Martinsen 2011; Moscardelli
692 and Wood 2015). As discussed here however, along active margins, shelfal environments not only
693 develop in the submerged prolongation of the main hinterland, constituting continent-attached shelves,
694 which can source large-scale shelf-derived mass-wasting products (e.g., Claussmann et al. 2021). Shelfal
695 environments can also form episodically upon and around structurally controlled and active ridges,
696 resulting in shelves that are detached from the main shelf domain (e.g., Ćosović et al. 2018). These
697 isolated settings, with a virtually absent to limited catchment area (whether underwater or emerged),
698 feature reduced shelfal source systems, which will, in turn, produce significantly smaller shelf-derived
699 mass-wasting products such as the ones described at Homewood. The size of the resulting deposits best
700 compares to those commonly found in *detached* MTDs (*sensu* Moscardelli and Wood 2015) (Appendix
701 6) whereas their internal characteristics inform that tectonic acted as the main triggering mechanism and
702 that the source was nearby.

703 Notwithstanding sea-level changes and high sedimentation rates as causal mechanisms for shelf-derived
704 mass-wasting events (Posamentier and Walker 2006; Moscardelli and Wood 2008; Moscardelli and
705 Wood 2015; and references within), Bailleul et al. (2013) and Claussmann et al. (2021) previously
706 demonstrated that along active margins, periods of tectonic activity (ca. 1 to 2 Ma) constitute another
707 mechanism to consider for both the development and subsequent shelf failures. The MTDs described in
708 this study not only provide another example of such shelf-derived mass-wasting products, whereby
709 tectonic activity controlled the repeated destabilization and collapses of the shelfal source (see also
710 Romero-Otero et al. 2010; Ortiz-Karpf et al. 2018), but also indicate that shelf-derived mass-wasting
711 events do not always result in large-scale systems (Appendix 6) (see also Claussmann et al. 2021).
712 Similarly, Naranjo-Vesga et al. (2020) observed that owing to the complex relationships that exist
713 between the shelf width, slope profile and sediment supply in an analogous convergent setting (*i.e.*,
714 southern Caribbean fold-and-thrust belt), both *attached* and *detached* MTDs could be sourced from the
715 regional shelf and slope regions.

716 Therefore, we here propose to slightly amend the MTD nomenclature from Moscardelli and Wood (2008)
717 and Moscardelli and Wood (2015) by using the related '*attached*' and '*detached*' morphometry
718 characteristics independently from the type (e.g., shelf, slope, local bathymetric highs) and distance from
719 source areas to better account for the variety of geotectonic settings (e.g., convergent).

720 As a result, we introduce the term '*shelf-derived* systems' which refers to all the MTDs that result from
721 the destabilization and collapses of shelfal environments, whether induced by regional and or local causal
722 mechanisms (sea-level changes, high sedimentation rates, periods of repeated tectonic activity, etc.),

723 and which can either source large-scale '*attached shelf-derived systems*' or smaller '*detached shelf-*
724 *derived systems*' (*sensu* Moscardelli and Wood 2008; Moscardelli and Wood 2015) (Figure 16). Similar
725 reasoning can be applied to the MTDs that result from failures of the regional slope (*i.e.*, '*slope-derived*
726 *systems*').

727 Notwithstanding the slope profile, here inherently linked to the geometry of the underlying structural styles
728 at the source (*e.g.*, forelimb or backlimb) and the sediment supply as other critical parameters (*e.g.*,
729 Naranjo-Vesga et al. 2020), the present study demonstrates that the geomorphology of source area, and
730 more particularly its connection to the continent (*i.e.*, catchment area) is also critical. Here, the narrow
731 shelves appear to source both '*attached and detached shelf-derived systems*' when connected to the
732 main hinterland (narrow continent-attached shelf) and '*detached shelf-derived systems*' when isolated
733 and disconnected from it (narrow continent-detached shelf) (Figure 16). Further work is however required
734 to generalize such an observation; and more particularly, the impact of the forelimb settings on the
735 resulting *shelf-* and *slope-derived* mass-wasting systems (*e.g.*, size, internal characteristics, causal
736 mechanisms) sourced from continent-detached mixed siliciclastic-carbonate shelfal domains remains to
737 be investigated.

738 8. CONCLUSIONS

739 In this study, we used high-resolution outcrop data to investigate the implications of mixed siliciclastic-
740 carbonate source systems developed upon and around the forelimbs and backlimbs of actively growing
741 asymmetrical structures for the development of the subsequent gravity-driven systems beyond the shelf
742 edges. Particularly, this study highlights the role and importance of varying structural setting of the
743 sediment source in controlling the morphologies of the delivery systems, the patterns of sediment
744 dispersal, and the related sedimentary facies along active margins.

745 First, we demonstrated that during the Middle Miocene, the paleogeography of the southern, inner portion
746 of the Hikurangi Margin was diversified, comprising a mature hinterland to the west as well as isolated
747 islands to the east, developing atop the actively growing seaward-verging thrusts of the subduction
748 wedge. Mixed compositional siliciclastic-carbonate sedimentation dominated at shallow depths and the
749 different morphologies of the source regions (continent-attached, forelimb and continent-detached,
750 backlimb) resulted in the development of highly varied shelf-derived gravity-driven systems downslope
751 within very short distances (less than five kilometers) that interacted with the structures and were
752 eventually captured both on their margins and within the trench-slope basins.

753 Then, we showed that mixed siliciclastic-carbonate gravity-driven systems in active settings appear
754 preferentially characterized by deposition of tens to a few hundred of meters of thick units, displaying
755 several kilometers of lateral extent, for duration of ca. 1 to 2 Ma. Such systems are primarily controlled
756 by the geometries and tectonic motion of the underlying structures at the shelf edges. Different
757 morphologies of the source regions (*i.e.*, the aforementioned continent-attached forelimb and continent-
758 detached backlimb of thrust settings) result in the development of distinct paleoenvironmental conditions
759 at the shelf and contrasting shelf-derived gravity-driven systems along and downslope. The shelf-edges
760 on the forelimbs of actively growing thrusts are abrupt and commonly associated with irregular steep
761 slopes. Here, they appear to have favored the development of systems that dynamically rejuvenate the
762 slopes (*e.g.*, Waikaraka retrogressive systems). On the backlimbs however, the shelf-edges are generally
763 associated to gentler slopes; yet, pulses in tectonic activity will increase slope gradient variations, thus
764 promoting the development of sedimentary systems that directly interact with the slope (*e.g.*, Homewood
765 mass-wasting system).

766 Finally, this study highlights that, whether they are caused by regional and or local triggering mechanisms,
767 the destabilization and collapses of shelfal environments source '*shelf-derived* mass-wasting systems',
768 characterized by two types: (1) large-scale, regional '*attached shelf-derived systems*' and (2) small-scale,

769 localized '*detached shelf-derived systems*'. Along active margins, and particularly the Hikurangi Margin,
770 the '*shelf-derived mass-wasting systems*' are sourced from shelves that are either attached to the main
771 continental domain or detached from it (e.g., isolated islands). The associated *shelf-derived* mass-
772 wasting deposits can be found on both sides of actively growing asymmetrical thrusts. When sourced
773 from the backlimbs however, the subsequent shelf-derived MTDs exhibit more complex internal
774 architectures, ultimately recording the dynamic changes in slope gradient, which can both trigger the
775 generation of mass failures and mass flows upslope as well as the coeval remobilization (translation and
776 deformation) of the previously deposited MTDs downslope. As a result, the MTDs from thrust backlimbs
777 can be used as proxies for unraveling the tectonic activity of an individual structure.

778 9. ACKNOWLEDGEMENTS

779 This research was funded by Schlumberger. We are grateful to Karen and John Barbour for their warm
780 welcome during our stays at Homewood and for access to the studied areas. Special thanks to Pierre
781 Malié and Andréa Barrier for their support during the fieldwork missions, as well as to Romain Armand,
782 François-Xavier Joanny and Pierre Saulet from UniLaSalle (France) for their technical assistance on the
783 GIS and outcrop modelling workflows. We also thank our colleagues from the Basins-Reservoirs-
784 Resources B2R research unit (University of UniLaSalle, France) for their lasting support and Hugh
785 Morgans and Henry Gard from GNS Science (Lower Hutt, New Zealand) for the paleontological analyses.
786 Finally, we thank the Editor Roberto Tinterri and the reviewers, Bradford Prather and Domenico Chiarella
787 for their helpful and constructive comments and suggestions.

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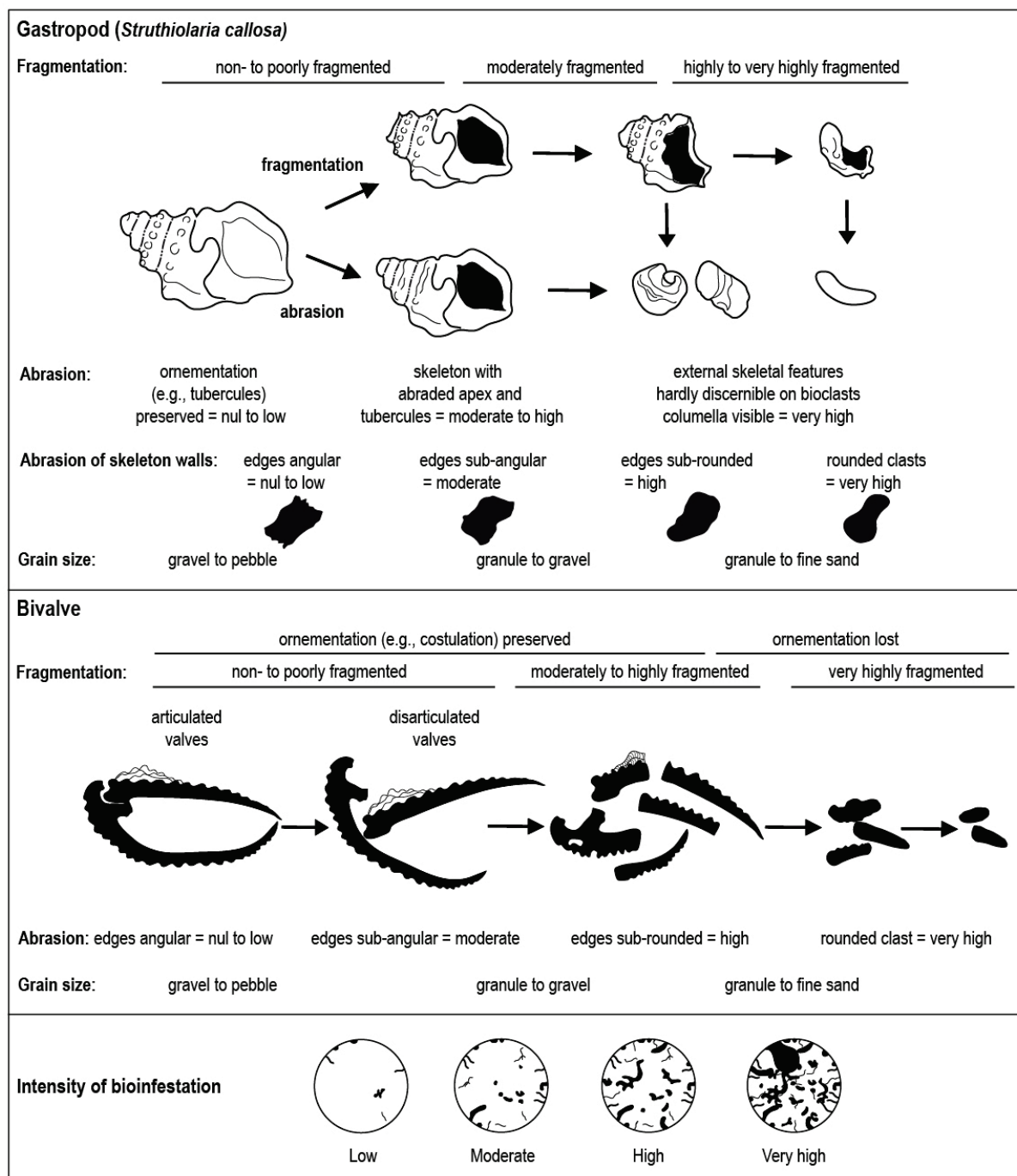
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11. SUPPLEMENTARY MATERIAL



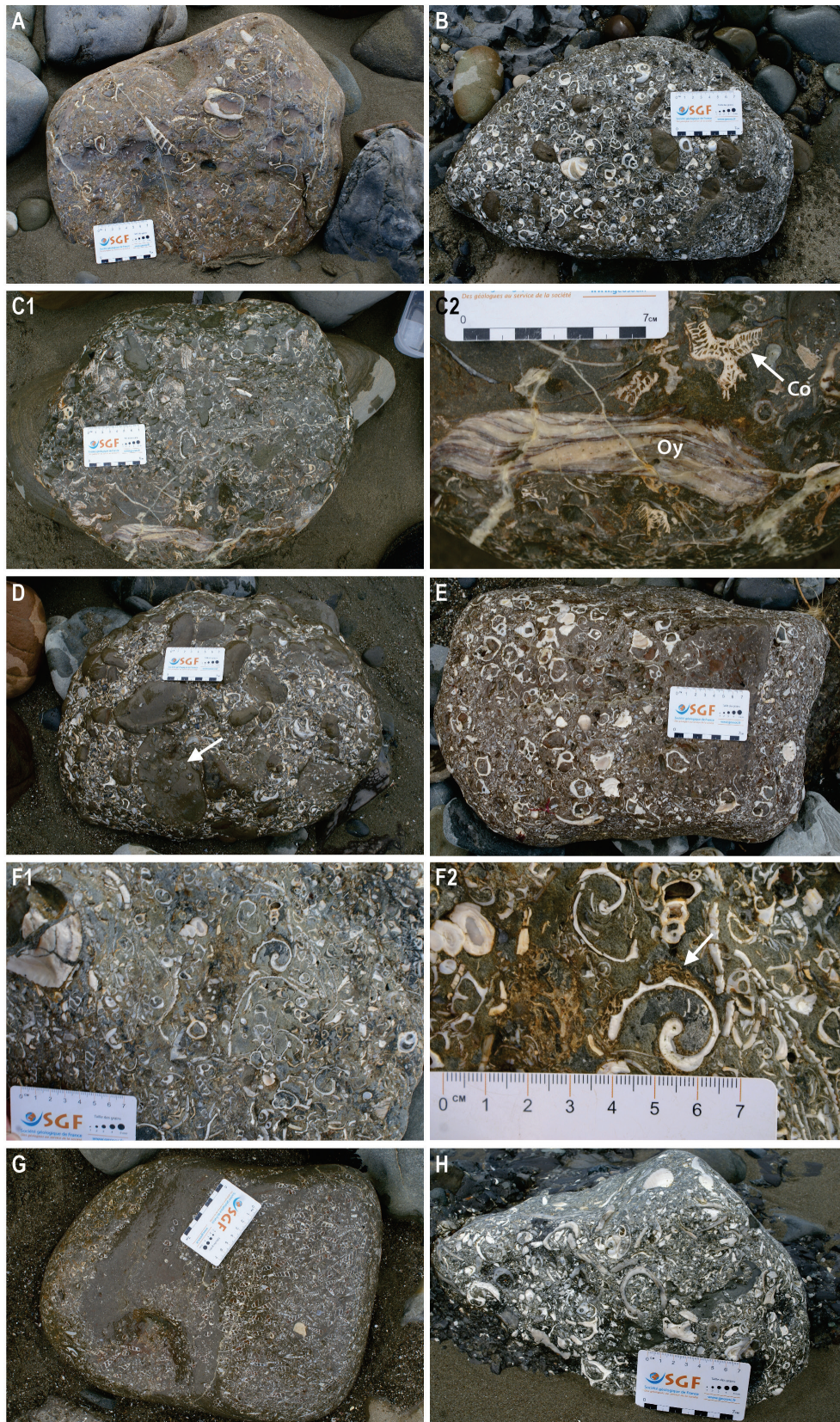
Appendix 1: Taphonomic methodology used on the macrofossils observed in the Waikaraka and Homewood localities. Qualitative taphonomic grading system established for the two dominant categories of fossil remains found in the studied deposits, namely Gastropods and Bivalves. The scales of alteration features were established to help characterize poorly to very highly damaged grains at outcrop. The particle contours displayed derive from field observations. The degree of abrasion was determined using the visual chart of (Powers 1953). The scale of skeleton bio-infestation is modified from (Perry 1998). The photographs provide field examples.

Sections	Fossil record Nb. & Field Nb.	LF code	Collector	Identifier	Date	Age	Age comment	Paleoecology & Paleobathymetry	Longitude	Latitude
	T27/0631 18_05_01	MDST	Clausmann, B.	Morgans, H.E.G. Gard, H.J.L.	02/2018	Upper Sc Lower Si	Planktic fauna suggests Late Cretaceous to Early Lillburnian (Middle Miocene; ca. 15.5 – ca. 14 Ma). No age diagnostic molluscs were present. Stratigraphically located above sample 18_05_06 and below T27/0493.	Deep middle bathyal (>800 m) or deeper. Planktic abundance >90% indicates a fully oceanic water mass overlay the site of deposition.	175,99163300	-41,19168700
Homewood-1 section	T27/0634 18_05_06	DF	Clausmann, B.	Morgans, H.E.G. Gard, H.J.L.	02/2018	SI - Tt	Broadest age is Lillburnian to Early T ongaporutuan (Middle to Late Miocene; Late Langhien - Tortonian; 15.1 - 7.2 Ma). The absence of <i>Paraglobobolalia mayeri</i> could indicate Early T ongaporutuan. Stratigraphically located below sample 18_05_01.	Middle bathyal paleodepth (>700 m). Although not abundant, the planktic fauna is 95% of the total foraminiferal fauna and indicates a fully oceanic water mass.	175,99264500	-41,18961700
	T27/0493	DF	Crundwell, M.P. Watson, J.	Crundwell, M.P.	01/1992	SI	Restricted to Lillburnian age (Middle Miocene; Late Langhien - Serravalian; 15.1 - 13.05 Ma). Stratigraphically located above sample 18_05_01 and 18_05_06.	Upper bathyal. Planktic abundance is 20%. Marginal oceanic slightly anoxic bottom conditions. Abundant Thalmammalia-like taxa (turbidite fauna?).	175,99248°E	41,1905°S
	T27/0625 18_03_02a	DF/CF	Clausmann, B. Maheux, G.	Morgans, H.E.G. Gard, H.J.L.	02/2018	SI - Tt	No planktics were recovered. The benthics indicate an age from Runangan to Waipian (Late Eocene to Pliocene; 34.6 - 3.00 Ma). Associated molluscs (<i>Struthiolaria callosa</i>) give a Lillburnian to T ongaporutuan age (Middle to Late Miocene; Late Langhien - Tortonian; 15.1 - 7.2 Ma).	The presence of <i>Cibicides molestus</i> indicates deep outer bathyal paleodepth (>150 m). The presence of molluscs indicates shallow waters.	176,03276700	-41,15014600
	T27/0626 18_03_02b	DF/CF	Clausmann, B. Maheux, G.	Morgans, H.E.G. Gard, H.J.L.	02/2018	Lw - Sw	Only benthic recovered has a range of Waikikian to Waiauan (Late Oligocene to Middle Miocene; Late Chattian - Serravalian; 25.2 - 11.04 Ma). No age diagnostic molluscs were present.	No real data. The molluscs indicate shallow waters and originate from mid to inner shelf environment.	176,03276700	-41,15014600
Waikaraka section	T27/0627 18_03_02c	DF/CF	Clausmann, B. Maheux, G.	Morgans, H.E.G. Gard, H.J.L.	02/2018	SI - Tt	A very poor foraminiferal fauna suggests Runangan to Recent (Late Eocene - Recent; 36.7 - 0 Ma). Molluscs give a Lillburnian to T ongaporutuan age (Middle to Late Miocene; Late Langhien - Tortonian; 15.1 - 7.2 Ma).	No real data. The molluscs indicate shallow waters and originate from mid to inner shelf environment.	176,03276700	-41,15014600
	T27/0628 18_03_02e	DF/CF	Clausmann, B. Maheux, G.	Morgans, H.E.G. Gard, H.J.L.	02/2018	NA	Only some possible agglutinated taxa and a nodosanid were observed.	Marine.	176,03276700	-41,15014600
	T27/0644 20_16_17	LDTc	Clausmann, B. Maheux, G.	Morgans, H.E.G.	12/2020	Sw	A good planktic fauna was recovered including a sinistrally coiled population of <i>Globobolalia mitulimida</i> , <i>Globoquadrina dehiscentis</i> and <i>Paraglobobolalia mayeri</i> . The age is Waiauan (Middle Miocene; Serravalian; 13.05 - 11.63).	The presence of <i>Tritaxilina zealandica</i> , <i>Sigmatolopsis schlumbergeri</i> , <i>Karriella cylindrica</i> indicates paleodepths at 1500 metres or deeper (lower bathyal). Abundant <i>Amphisigina</i> , and less common <i>Zenitellus</i> , <i>Florilus</i> , <i>Ehrenbergina</i> , and <i>Cibicides perforatus</i> indicates shallow to middle shelf reworking across the shelf into deeper water. Planktic abundance is about 40% indicating a middle neritic watermass.	176,02902333	-41,15342233

Appendix 2: Sample list used for the Whareama Basin outcrops of Homewood and Waikaraka. The Fossil Record numbers including "T27" relate to the T27 New Zealand Map Grid 1:50,000 series and are captured in the Fossil Record Electronic Database (<https://fred.org.nz>). See Table 1 for the description of the associated lithofacies code. The age of the sedimentary units and the paleobathymetries were defined using micro- and macropaleontological analyses conducted by GNS Science New Zealand.



Appendix 4: Enlarged view of the Homewood outcrop, mostly comprising a several hundreds of meter-thick distal depositional lobe system (Fa1g). This study particularly focuses on the gravity-driven systems present at the base of this series and which are presented in the detailed view of the outcrop model in Figure 9.



Appendix 5: Clasts of the inner- to mid-shelf shell beds clasts encountered at Waikaraka. (A): Siliciclastic floatstone dominated by gastropods (*Struthiolaria* sp. and turrillids) and thick-shelled bivalves in a sandstone matrix. (B): Pebbly molluscan rudstone. Sandstone clasts include sub-rounded to rounded gravels and pebbles. (C1-2): High-diversity clast-supported

rudstone containing thick-shelled oysters (Oy, C2), zooxanthellate branching corals (Co, C2), brachiopods, turrnellids, serpulids and subsidiary bryozoans. (D): Poorly-sorted pebbly rudstone containing abundant bored, well-rounded, sandy pebbles, and highly fragmented and abraded molluscan remains. (E): Molluscan floatstone-rudstone rich in thick-shelled gastropods and bivalves. (F1-2): Molluscan rudstone includes both thin- and thick-shelled molluscs commonly encrusted by coralline red algae (arrow, F2). (G): Turrnellid floatstone. (H): Densely packed, thick-shelled molluscan rudstone.



Calculations using Moscardelli & Wood 2015

GENERAL

Equation (1): $L = 1.5051 * (A^{0.4977})$

Equation (4): $V = 0.0008 * (T^{2.1998})$

Equation (12): $V = 0.0281 * (A^{1.1123})$

Rearranged Equation (12): $A = ((V) * (1/0.0281))^{(1/1.1123)}$

DETACHED

Equation (2): $Ld = 1.2477 * (Ad^{0.5415})$

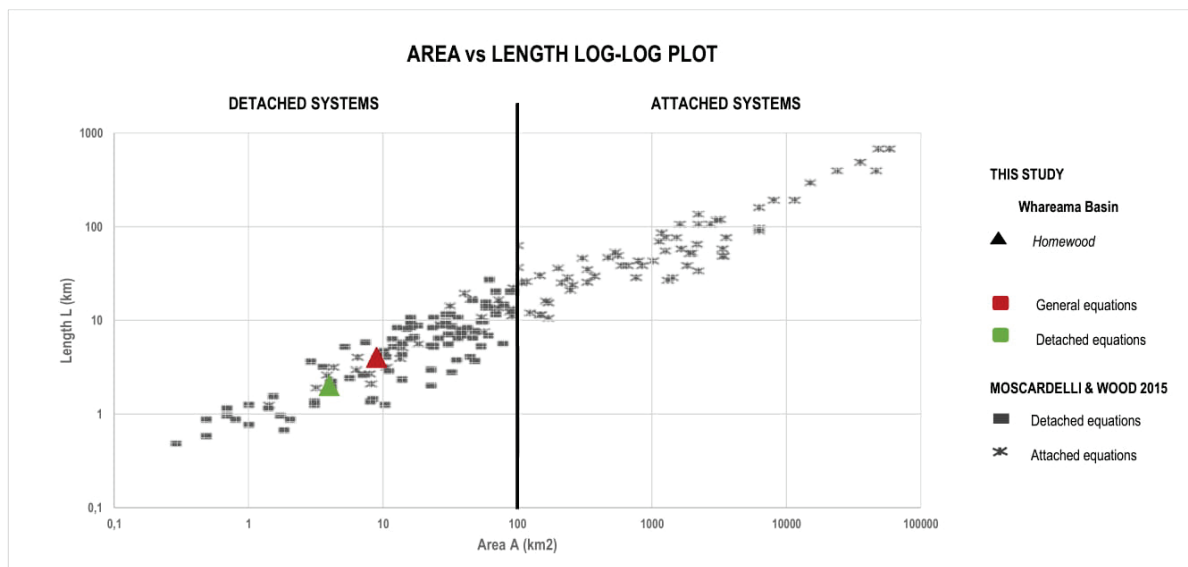
Equation (6): $Vd = 0.0014 * (Td^{1.6223})$

Equation (14): $Vd = 0.0261 * (Ad^{1.1507})$

Rearranged Equation (14): $Ad = ((Vd) * (1/0.0261))^{(1/1.1507)}$

USING OUTCROP THICKNESS

		Thickness T (m)	Volume V (km3)	Area A (km2)	Length L (km)	Equations
WHAREAMA	INPUT	Sefton Hills-1				
		15				
						RESULT
			Equation (4)	Rearranged Equation (12)	Equation (1)	General
			0,31	9	4	
			Equation (6)	Rearranged Equation (14)	Equation (2)	Detached s.l.
			0,11	4	2	



Appendix 6: Morphometric parameter calculations using the equations from (Moscardelli and Wood 2015) for the Homewood mass-transport deposits. The outcrop conditions prevented the recognition of most of the morphometric parameters and therefore the calculations were estimated using the assumption of a relative uniform thickness of 15 meters. The calculations were done using the general set of equations as well as the dedicated set for “detached systems”.