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Response of the Rhône deltaic margin to loading and subsidence during the last climatic cycle

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Abstract:

Passive continental margin subsidence is initiated by the synrift mechanical stretching of the lithospheric upper brittle layer and continues during the postrift phase; the thermal cooling and contraction of the upwelled asthenosphere forces the margin to subside in addition to the overloads from sea water and sediments. Therefore, the total subsidence in stretched basins includes fault-controlled initial sinking, thermal subsidence and flexural isostatic compensations. Decoupling and estimating the different components of this subsidence from stratigraphic analysis and restricted geophysical and sedimentological databases remains problematic. In particular, backstripping the sediment layers requires a well-constrained geological framework. A method is proposed here to investigate the subsidence history of a margin based on forward stratigraphic modelling. Using the Sedflux model, several experiments are done using generally agreed upon assumptions on the parameters describing lithospheric rheology and isostatic behaviour of a margin. The stratigraphic modelling of the Rhône deltaic margin during the last climatic cycle (125 kyr) provides an assessment of these parameter estimates and their influence on geohistory (tectonic/thermal subsidence and sediment loading). The model results confirm the important impact of water loading on vertical deflection along the platform between glacial low sea-level and interglacial high sea-level. Based on Gulf of Lions (NW Mediterranean) observations, a conceptual method that uses the stratigraphic simulations is produced in order to evaluate the different components of the total subsidence of a margin, and, in particular, the relative impact of tectonic subsidence and sediment load.

Keywords: Subsidence; Isostasy; Stratigraphic simulations; Sedflux; Continental shelf; Gulf of Lions

1. Introduction

Present-day stratigraphic organisation and sedimentary thickness on a platform are products of cumulative changes in sedimentary systems through time. The location and preservation of depocentres, as shown by seismic profiles from the shelf, result from changes in accommodation. A significant subsidence rate of the margin is necessary to permit a continuous record and preservation of a depositional sequence. On passive margins,

55 accommodation is most important at the shelf edge. The location and magnitude of the
56 sediment sources, together with eustatic controls, may result in erosion of the sediment wedge
57 on the inner shelf. Studying basin subsidence and sedimentary filling is essential for
58 understanding the tectonic and thermal history of passive continental margins. Therefore,
59 subsidence rate, together with sediment flux and global sea-level variations have to be taken
60 into account in order to investigate the origin of vertical motions of marine continental
61 shelves. Total subsidence, which contributes to accommodation, corresponds to medium to
62 long-term Earth processes that involve constraints from lithospheric structure and
63 asthenospheric cooling. The overloads of sediments and water amplify vertical movements,
64 according to the laws of isostatic compensation. Numerous parameters are implied for this
65 process. It is therefore difficult to disentangle and quantify the different components that
66 contribute to the subsidence of the platform.

67

68 Geohistory analysis (Van Hinte, 1978), based on seismic stratigraphy and lithological
69 information from boreholes or sediments cores, provides important constraints on the tectonic
70 and/or thermal subsidence and sediment accumulation rates through time. Decompaction of
71 the present-day sedimentary thicknesses, paleobathymetry and paleosealevel helps to evaluate
72 the vertical evolution of a continental margin. Using this method, amounts of total and
73 tectonic subsidence can be determined through decompaction of stratigraphy and
74 backstripping of sediment load. This technique has been applied to investigate several
75 margins worldwide (Ceramicola *et al.*, 2005, Steckler *et al.*, 1999). Watts and Ryan (1976)
76 proposed the isolation of the tectonic driving force by removal of the isostatic effects of
77 sediment load (i.e. the Backstripping technique). Unfortunately, the validity of the method
78 requires a well-known sedimentary and structural system and precise estimates of specific
79 parameters like compaction, paleobathymetry and absolute sea-level fluctuations. A

80 quantitative analysis of subsidence rate through time relies on knowledge of basin formation
81 and evolution.

82

83 There are two conceptual approaches to model a basin and to determine the factors that affect
84 its formation and its infilling. The first is the backstripping method, which successively
85 removes strata to recover a margin's geohistory. Most of the time, it is not simple to estimate
86 each of the parameters required by this method. This leads to under- or over-estimation of the
87 subsidence components. In particular, determination of paleobathymetry from observational
88 data cannot generally be estimated with enough accuracy. In addition, the sediment record is
89 discontinuous because of non-deposition and erosional events, and the complete geodynamic
90 evolution of margins cannot be reconstructed (i.e. Bessis, 1986; Ceramicola *et al.*, 2005,
91 Steckler *et al.*, 1999). Thus, backstripping implies assumptions about the geological evolution
92 of the studied margin, without possibility for testing the validity of these hypotheses. The
93 second approach, described here, uses forward stratigraphic modelling to simulate the
94 delivery of sediment and its accumulation in a sedimentary basin (e.g. Syvitski and Hutton,
95 2001), and takes into account variations in the various controls on sedimentation. In
96 conjunction with the backstripping method, the stratigraphic simulation is useful for
97 investigating the boundary conditions of the evolution of a basin and its changes in
98 accommodation.

99

100 This paper demonstrates the use of stratigraphic simulations to validate and extrapolate
101 different hypotheses on basin dynamics and the formation of shelf sedimentary wedges. An
102 estimate of subsidence and sediment thickness from seismic stratigraphic analysis is tested
103 and refined during a simulation of the basin evolution. The method is applied to the Gulf of
104 Lions as a case study. Next, a conceptual method using the stratigraphic simulations is

105 produced to isolate the different components of total subsidence of a margin (tectonic/thermal
106 subsidence, sediment and water loading) during a relatively short geological interval.

107

108 **2. Subsidence and Isostasy**

109

110 Sediment dispersion and deposition in a sedimentary basin is the product of the interplay
111 between the generation of accommodation and sediment supply. Sediment accumulations and
112 their internal geometries are therefore controlled largely by the tectonic/thermal and isostatic
113 mechanisms that cause subsidence (Fig. 1). Change in accommodation is thus an important
114 part of the driving mechanisms responsible for the stratigraphic pattern in basin-fill. If we
115 consider a theoretical lithological column, these mechanisms can be subdivided into: 1-
116 tectonic forcing controlling the spatial and temporal pattern of subsidence and the evolution
117 of the sediment routing system, and 2- eustasy that essentially controls accommodation and
118 sets base level (Allen and Allen, 2005). As a consequence, several vertical motions of the
119 reference level can be observed (Fig. 1). Motions can be inferred from the evolution of each
120 lithological unit and lead us to define the different components of the total subsidence (*S*).
121 The lithospheric structural processes of margin formation and evolution through time are
122 involved in the tectonic/thermal subsidence (*TS*), whereas global eustasy and sedimentation
123 are associated and have an important impact on the isostatic response of this margin. Sea
124 water defines the water loading (*WL*), which relies on sea-level variations. The thickness of
125 sedimentary accretion is the result of erosion/deposition processes; it contributes to the
126 sediment load (*SL*). We define "geohistory" subsidence (*GS*) as the combination of *TS* and *SL*.
127 Seismic stratigraphy is one way to evaluate *GS* through time, often between two successive
128 interglacial high sea-levels in order to remove the effects of sea-level variations. Below we

129 discuss each component of the total subsidence and, in particular, the proportion of tectonic
130 subsidence and sediment load in the geohistory subsidence (Fig. 2).

131

132 ***Insert Figure 1***

133

134 **2.1 Tectonic/thermal subsidence: basin formation**

135

136 Our study focuses on the shelf area of passive continental margins (i.e. seismically inactive).
137 In a uniform stretching model (McKenzie, 1978), the formation of a passive margin can be
138 divided into two major phases of structural adjustments. During continental rifting, there is a
139 brittle extension of the crust that produces the stretching of the continental lithosphere and a
140 rapid synrift subsidence. After lithospheric thinning, a postrift phase is mainly governed by
141 the cooling and contraction of the upwelled asthenosphere. The thermal relaxation is at the
142 origin of an exponentially decreasing postrift subsidence. Synrift tectonic subsidence rates are
143 typically <0.2 mm/y (200 m/My) and postrift tectonic subsidence rates are about <0.05 mm/y
144 (50 m/My) (Allen and Allen, 2005). The amount of synrift and postrift subsidence depends
145 essentially on the initial crustal to lithospheric thickness ratio and on the amount of stretching.
146 The present-day tectonic/thermal subsidence of a margin is a long-term geological process
147 that directly results from its structural and thermal context and history, and from its age
148 (Fig. 1).

149

150 **2.2 Isostatic subsidence**

151

152 Increases in sediment load and water load causes additional subsidence of the sedimentary
153 basin through the isostatic response of the lithosphere.

154

155 **2.2.1 Subsidence from overload**

156

157 Passive margins are characterized by seaward thickening prisms of marine sediments
158 overlying a faulted basement of synrift sedimentary sequences. The postrift seaward-
159 thickening sediment prisms consist predominantly of shallower marine deposits (Allen and
160 Allen, 2005). This sedimentary wedge, together with the overlying sea water column modifies
161 long-term tectonic subsidence. The weight of sediment deposited on a particular area of the
162 shelf may cause the underlying crust to sink. Sediment erosion may cause the margin to rise.

163

164 A change in global sea-level relative to a reference datum is known as eustasy (Lisitzin,
165 1974). Eustasy in turn is one of the major causes of relative sea-level changes through hydro-
166 isostasy (Johnston, 1995; Lambeck, 1997, 2000; Peltier, 2002; Posamentier *et al.*, 1988), and
167 thus impacts accommodation. Any increase (or decrease) of ocean volume must be
168 compensated isostatically. Global sea-level changes are largely due to global climatic
169 changes. As the earth's climate cools, the ocean surface cools and ocean volume decreases
170 (the steric effect). Additionally ice-sheets may form, storing water on land and reducing the
171 ocean volume. During a warming period, ocean volume changes will move in the opposite
172 direction. The major consequences of sea-level rise are observed at the glacial/interglacial
173 transition, with differences between low and high sea-level positions of more than 120 m;
174 between successive interglacial sea-level positions, the differences are minor.

175

176 **2.2.2 Flexure of the lithosphere: isostatic compensation**

177

178 Below some depth, there is no density contrast between two adjacent columns. The weight of
179 the columns above this depth of compensation must be equal. This is a local isostatic balance
180 (Airy, 1855; Pratt, 1855); the deflection of the crust at any location depends only on the local
181 overload at that location (Airy and Pratt models of compensation). But the local isostatic
182 balance neglects the lateral strength of the lithosphere and its relative rigidity. A more
183 realistic model assumes that the lithosphere responds to loads like an elastic plate overlying
184 an inviscid fluid (Kirby, 1983). Application of Archimedes principle suggests that bent
185 continental plates are buoyed up by a force equal to the weight of the displaced mantle
186 (Turcotte and Schubert, 1982). The net effect is for the entire region affected by flexure to be
187 in regional isostatic balance. The lithosphere behaves approximately as an elastic beam of
188 some assumed rigidity. A more rigid beam produces a broader and shallower deflection. A
189 less rigid beam results in a deeper and narrower deflection.

190

191 A quantitative way to estimate the rigidity of the lithosphere is its effective elastic thickness
192 (*EET*) (Burov and Diament, 1995; Watts, 1992).

$$193 \quad EET = \sqrt[3]{\frac{12(1-\nu^2)D}{E}} \quad (1)$$

194 Two constants, the Poisson's ratio ($\nu = 0.25$) and the Young's modulus ($E = 7.10^{10} \text{ N/m}^2$),
195 characterize the rheology (the stress/strain relationship) of the elastic portion of the crust and
196 mantle lithosphere. D is the flexural rigidity in N/m. *EET* appears to be independent of the age
197 of the load (Watts *et al.*, 1982), which suggests that the elastic stresses that cause deflection,
198 do not relax on a geological time scale. Adding a sediment and/or water load to the deflection
199 causes the amount of deflection to increase. The most complete dataset currently available
200 (Watts, 2001) describes values of *EET* between 5 and 110 km. For the specific application in
201 the Gulf of Lions, Lambeck and Bard (2000) use an *EET* of 65 km in order to model the last
202 deglacial isostatic rebound.

203

204 Using a simple model for bending a visco-elastic slab under a distributed load permits us to
205 explore the effects of the isostatic components within a single-layer lithosphere. In the case of
206 a glaciated margin, Huybrechts and De Wolde (1999) propose an isostatic ice-dynamic
207 reconstruction model that assumes a rigid elastic lithosphere overlies a viscous asthenosphere.
208 In this model bedrock adjustments are described by a single isostatic relaxation time. In this
209 way, the isostatic compensation takes into account the effects of loading changes within an
210 area several hundred kilometres wide. From this example, we obtain a value for the flexural
211 rigidity (10^{25} Nm) corresponding to a lithospheric thickness of 115 km; the characteristic
212 relaxation time for the asthenosphere is about 3,000 years (Huybrechts, 2002). The relaxation
213 time is characteristic of the time dependence of the isostatic rebound process; it depends
214 almost entirely upon the viscosity of the mantle. Relaxation times vary from approximately
215 3,400 years in SE Hudson Bay (Canada) to 4,200 years in the Gulf of Bothnia (Sweden) for
216 postglacial rebound modelling (Peltier, 1998).

217

218 Thus, investigating the present-day stratigraphy should help us to estimate the
219 accommodation history and evaluate the different components of subsidence that contribute to
220 its changes. The typical response of continental stretching is early, rapid, fault-controlled
221 subsidence followed by lithosphere cooling dominated by gravity-controlled deformation.
222 Sediment accumulation and water load in a sedimentary basin causes extra-subsidence of the
223 basement corresponding to a general basinward tilt in a long-term postrift tectonic and
224 thermal subsiding context. The isostatic flexure of the lithosphere undergoing additional
225 sedimentary load, such as a prograding sedimentary wedge, produces a regional deflection
226 that is controlled by the effective elastic thickness of the lithosphere and the properties of the
227 mantle viscosity (modelled by the relaxation time parameter). However, if the lithosphere

228 reacts to the sediment load through regional flexure, the separation of the tectonic and
229 sediment contributions is complex. The flexural loading of the sedimentary basin can be
230 accounted for if the flexural rigidity and spatial distribution of the sediment load is known.

231

232 **3. Methods for estimating subsidence**

233

234 This study defines a method to better constrain the environmental and structural settings of a
235 study area (Fig. 2), especially the isostatic behaviour of a continental shelf under different
236 loading conditions (variations in sediment and water loads). There are different ways to
237 estimate components of subsidence and to approach a margin's isostatic characteristics. The
238 weakness of methods, such as "backstripping", is that they often require much knowledge
239 about general settings that lead to many assumptions. In order to fix the isostatic parameters,
240 one solution is to test them in their context and take advantage of stratigraphic simulations.
241 Different hypothesis can be tested. Numerical stratigraphic models are useful for
242 understanding the time-varying impact of sedimentary processes on the stratigraphic
243 organisation. Stratigraphic simulation models are based on algorithms that conceptually or
244 dynamically simulate the important input, boundary conditions and processes that define a
245 sedimentary system (Syvitski, 1989). Subsidence, sea-level and isostasy combine to create
246 accommodation in the basin, which controls sedimentation on the shelf. They also correspond
247 to the main input values required for the simulations. In this way the different components of
248 subsidence, measured during geohistory analysis, can be tested and estimated through
249 numerical modelling.

250

251 ***Insert Figure 2***

252

253 **3.1 *Sedflux*: eustasy and subsidence modelling**

254

255 Here, we apply the model *Sedflux* (Hutton & Syvitski, *this volume*) to simulate the delivery
256 and accumulation of sediment within a basin through time. *Sedflux* includes the effects of sea-
257 level and sediment supply fluctuations over time scales of tens of thousands of years. The
258 basin-fill model allows for the continental margin to undergo tectonic processes (subsidence
259 and uplift) and isostatic effects from sediment and water loads. The model architecture has a
260 typical vertical resolution of 1 to 25 cm, and a typical horizontal resolution of 10 to 100 m.
261 Various processes are modelled at a time step (days to years) that is sensitive to median-term
262 variations of the seafloor (Syvitski and Hutton, 2001). A major subroutine of *Sedflux*
263 corresponds to a momentum-driven hypopycnal plume, based on the Albertson *et al.* (1950)
264 model of a submerged and steady, two-dimensional surface jet emanating out of a river mouth
265 (Syvitski *et al.*, 1998). This advection-diffusion subroutine introduces a time-varying
266 sediment flux into the modelled basin to allow the stratigraphic organisation of sediment on a
267 shelf. *Sedflux* requires as inputs an initial bathymetry of the basin at the simulation onset, and
268 time-varying sediment flux and sea-level history. Of importance to this study are parameters
269 related to subsidence and flexural response. In *Sedflux*, two different types of subsidence are
270 considered: isostasy and tectonic subsidence. For isostatic subsidence, the lithosphere is
271 treated as an elastic beam that is allowed to flex under the load of added sediment and water.
272 For tectonic subsidence the user specifies subsidence rates at various positions and times
273 (Syvitski and Hutton, 2001).

274

275 **3.1.1 Tectonic subsidence**

276

277 The physics of the processes that lead to tectonic or thermal movements are not modelled in
278 *Sedflux*. Instead the results of these processes are incorporated as input to the model domain;
279 vertical displacements for the modelled basin are specified in an input file, and these are
280 allowed to vary both spatially and temporally. Subsidence and uplift rates are defined in
281 meters per year at particular point along a basin for a specific instant in time. *Sedflux*
282 interpolates these data to the defined temporal and spatial resolution of the particular model
283 run. Results of the modelling provide confirmation of the range of subsidence rates used for
284 it.

285

286 **3.1.2 Isostatic subsidence**

287

288 The changes in water and sediment load in a basin cause vertical lithospheric deflections
289 (Fig. 1). In the case of a thick sedimentary wedge, subsidence becomes a leading process.
290 *Sedflux* models subsidence due to loading using an elastic flexure model (Syvitski and Hutton,
291 2001). The elastic flexure model applied to Earth's crust makes four basic assumptions; 1- the
292 lithosphere is assumed to have a linear elastic rheology, 2- the deflections are assumed to be
293 small, 3- the elastic lithosphere is assumed to be thin compared to the horizontal dimensions
294 of the plate, 4- planar sections within the plate are assumed to remain planar after deflection.
295 For a single vertical load applied to the Earth's crust, the resulting displacements are given
296 by:

$$297 \quad w(x) = \frac{p(x)\alpha^3}{8D} \exp\left(-\frac{|x|}{\alpha}\right) \left(\cos\left(\frac{|x|}{\alpha}\right) + \sin\left(\frac{|x|}{\alpha}\right) \right) \quad (2)$$

298 with α defined as

$$299 \quad \alpha \equiv \sqrt[4]{\frac{4D}{\rho_m g}} \quad (3)$$

300 and w is the displacement of crust due to sediment loading, D the flexural rigidity of the
301 Earth's crust (i.e. Equ.1), ρ_m the density of the overlying sediment, and x the horizontal
302 position. Because of our assumption of the linearity of our system, the resulting displacement
303 due to multiple columns of sediment is simply the sum of the displacements due to each
304 individual column.

305

306 After a load is applied, the viscous asthenosphere must flow out of the way before the
307 lithosphere can deflect; causing a time delay between the addition of load and the
308 lithosphere's response. Although models exist that predict the crustal response given a series
309 of viscosity layers (Paulson *et al.*, 2005), this is beyond the scope of *Sedflux*. Instead, *Sedflux*
310 assumes that the crustal response is exponential with time,

$$311 \quad w(t) = w_0 \left(1 - \exp\left(-\frac{t}{t_0}\right) \right) \quad (4)$$

312 where w_0 is the equilibrium deflection as determined by Equation (2), t is the time since the
313 load was applied, and t_0 is the response time of the lithosphere. The elastic flexure model in
314 *Sedflux* only needs values for the effective elastic thickness (Equ.1) and the relaxation time
315 (Equ.4) in order to calculate lithospheric deflections.

316

317 **3.2 Method strategy**

318

319 Based on the results from geohistory analysis (1 in Fig. 2), different parameters are used for
320 several numerical runs of the stratigraphic model to estimate the different components of the
321 total subsidence (S). The interpretation of seismic and lithological data allows the definition
322 of the geohistory subsidence (GS) (2 in Fig. 2). The identification of dated erosion
323 paleosurfaces permits us to quantify their vertical evolution through time and to estimate the

324 *GS* subsidence. This value takes into account both the tectonic subsidence (*TS*) and the loads
325 due to sediment deposition (*SL*). Using the *GS* estimation and the sea-level variations as input
326 parameters, the stratigraphic modelling can be realized with *Sedflux* (3 in Fig. 2).

327

328 The first stage of modelling is to experiment with different ranges of parameters that set the
329 isostatic adjustment, in order to define of the best effective elastic thickness and relaxation
330 time. The first model is run only with the *GS* subsidence (4 in Fig. 2), and provides a view of
331 the tectonic subsidence added to the sediment load effect on the margin during the simulated
332 time. The second model run does the same with eustasy (5 in Fig. 2). Therefore the margin
333 responds to changing water load due to sea-level variations and consequently water column
334 thickness fluctuates (*WL*). The combination of *GS* and eustasy is used for the third simulation
335 (6 in Fig. 2), and corresponds to the modelling of the total subsidence (*S*) as all components
336 are considered. These results are compared to the stratigraphic pattern observed on seismic
337 profiles and parameters are adjusted in order to minimize the differences between model
338 results and field observations (7 in Fig. 2).

339

340 The final stage of this method is to quantify the fraction of tectonic subsidence (*TS*) relative to
341 the sediment load (*SL*) within the *GS* subsidence. For this simulation, a constant sediment flux
342 is added to the *GS* subsidence and the eustasy, with the objective to reproduce the
343 sedimentary thicknesses observed on seismic analysis (8 in Fig. 2). We call the resulting
344 simulated subsidence *S'*, as it corresponds to the addition of total subsidence (*S*) with the
345 sediment load (*SL*) (9 in Fig. 2). The comparison between *S* and *S'* allows us to estimate the
346 effect of sediment load (*SL*) (10 in Fig. 2). Finally, the values of *SL* serve to partition *GS*
347 subsidence into sediment load (*SL*) and tectonic/thermal subsidence (*TS*). Therefore, the suite

348 of stratigraphic simulation provides an estimate of all the components controlling the vertical
349 motion of the margin.

350

351 **4. Application: Stratigraphic modelling of the Gulf of Lions margin**

352

353 The subsidence history of the Gulf of Lions margin (Fig. 3) is investigated using stratigraphic
354 simulations to calibrate our subsidence study. The available geophysical and lithological
355 datasets provide an ideal well-constrained domain of application. Subsidence was historically
356 explored on the basis of petroleum boreholes (Watts and Ryan, 1976) and multichannel
357 seismic (Bessis, 1986; Bessis and Burrus, 1986). More recently, subsidence rate was
358 established from high resolution seismic, constrained by modelling of the last 500 ky
359 (Rabineau, 2001). Our method, using the *Sedflux* model, has been assessed for the last
360 climatic cycle, and subsidence components have been estimated.

361

362 **Insert Figure 3**

363

364 **4.1 Geological settings**

365

366 The Palaeozoic and Mesozoic basement of the Gulf of Lions continental margin has
367 undergone several phases of stretches and strains since the Hercynian orogeneses (Biju-
368 Duval, 1984). The passive margin was shaped following the combined Oligo-Aquitania
369 rifting phase between continent and the Corsica-Sardinia microplate, and the Burdigalian
370 crustal opening (Gueguen, 1995; Sioni, 1997). This margin is covered by sedimentary series
371 dated from Oligocene to Quaternary (Bentounsi, 1990; Gorini *et al.*, 1993). The synrift series
372 (30-24 My) is topped by a Middle Aquitanian to Middle Burdigalian ravinement surface,

373 which marked the onset of clastic postrift deposits. This depositional sequence (24-6.3 My)
374 corresponds to the Miocene prograding wedge, largely eroded on the shelf and upper slope
375 during the Messinian crisis (6.3-5.2 My) (Lofi *et al.*, 2003). The upper Plio-Quaternary
376 deposits have recorded the sedimentary structures associated with the increasing sea-level
377 fluctuations during that time (Berné *et al.*, 2002).

378

379 The Quaternary stratigraphic organisation of the Gulf of Lions is described by several
380 conceptual models (Aloisi, 1986; Got, 1973; Monaco, 1971). High-resolution seismic data
381 show, within the Middle and Late Quaternary, the repetition of several prograding wedges
382 bounded by high amplitude seismic discontinuities (Fig. 4). These surfaces pinch out
383 landward at about 80 m water depth. Within each seismic sequence, two major types of
384 seismic facies are identified. Gently dipping clinoforms (PI) were interpreted as the product of
385 mud deposition in a relatively low-energy environment whereas relatively high-angle
386 clinoforms (PII, from 3° to 7°) were considered as corresponding to sandy upper shoreface
387 facies (Berné *et al.*, 1998; Gensous and Tesson, 1996; Rabineau, 2001; Rabineau *et al.* 2005;
388 Tesson *et al.*, 1990; Tesson *et al.*, 2000). The major shelf sequences are associated with the
389 Middle and Late Quaternary glacial/interglacial climatic and eustatic fluctuations. Regressive
390 deposits constitute the majority of preserved sediments. Using stratigraphic modelling,
391 Rabineau (2001) demonstrated that these sequences are linked to 100 ky orbital cycles.

392

393 **Insert Figure 4**

394

395 The present-day bathymetric configuration (Fig. 3) of the Gulf of Lions illustrates the present
396 highstand situation with distinct lowstand, forced regressed and highstand systems tracts. The
397 Holocene Rhône prodeltaic lobes and the last transgressive units, form the post-glacial

398 subaqueous delta on the inner shelf, and sediment accumulations along the coast. From the
399 middle to outer shelf, the majority of the prograding wedges correspond to regressive
400 deposits. Parts of the wedges consist of muddy sedimentary bodies with gently dipping
401 clinoforms while others form sandy shorefaces with high-angle clinoforms that settle on the
402 outer shelf (Berné *et al.*, 1998) (Fig. 5). This major seismic sequence, formed as a forced
403 regression during the overall sea-level fall between MIS-3 and MIS-2, corresponds to a falling
404 stage systems tract in the sense of Plint and Nummedal (2000). It can be sub-divided into
405 several prograding units, which indicate that this relative sea-level fall was punctuated by
406 intervals of increased or decreased falls, or even stillstand (Jouet *et al.*, 2006). Major
407 polygenetic regional erosion surfaces top the last two glacial sedimentary prisms. They
408 formed both as subaerial and marine erosion surfaces during sea-level fall (sequence
409 boundaries), and then were reworked (as a ravinement surface) during the ensuing
410 transgression (Bassetti *et al.*, 2006).

411

412 **Insert Figure 5**

413

414 The sequence deposited during the last climatic cycle is shown by a very-high resolution
415 Chirp seismic profile (Fig. 5). It is located between the erosion surface *D60*, attributed to the
416 penultimate glacial period (Marine Isotopic Stage 6 or MIS-6), and the erosion surface *D70*,
417 formed during MIS-2 (Fig. 6). During glacial periods, sea-level was at a relatively low
418 position, and favoured erosion. Although *D70* is defined as the last glacial erosion surface, we
419 can observe it within an interglacial (highstand of sea-level, MIS-1) situation. Later in this
420 paper, we will compare the present-day position (MIS-1) of *D70* to the position of *D60* (dated
421 to MIS-6) at its subsequent interglacial position (MIS-5). For each case, it corresponds to the
422 position of a glacial erosional surface at the following interglacial highstand.

423

424 **4.2 Subsidence in the Gulf of Lions**

425

426 During the Late Quaternary, the Gulf of Lions margin underwent postrift deformations. The
427 amount of well-preserved sediment accumulation on the shelf attests to a considerable
428 geohistory subsidence, which is the result of the combination of thermal subsidence and
429 sediment load effects. The quantitative estimation of the geohistory subsidence in the Gulf of
430 Lions was typically realised by the "Backstripping" method on depth converted seismic
431 sections crossing the margin, and using several petroleum exploration boreholes (Bessis,
432 1986; Watts and Ryan, 1976). The variations of subsidence from the internal platform to the
433 deep basin confirm the rapid initial burying of the margin between 30 to 23 My associated
434 with crustal stretching during rifting (mechanic tectonic subsidence). The curves illustrate the
435 exponential slowdown in postrift subsidence rate in response to the cooling of the lithosphere
436 (thermal tectonic subsidence) without any significant tectonic activity. The cumulative
437 subsidence of the basin (reaching 10 km in the deep basin) since the Oligocene would be
438 equivalent to that calculated for older Atlantic margins (Bessis and Burrus, 1986; Burrus,
439 1984) although its age is only 30 My. Such magnitudes cannot be explained by an extensional
440 model alone. The Gulf of Lions margin has the physiography of an Atlantic-type margin with
441 the subsidence rate of an active margin (100-200 m/My) (Steckler and Watts, 1980). For the
442 Upper Quaternary, Rabineau *et al.* (2005) estimates the geohistory subsidence rate at around
443 255 m/My at the shelf edge from the stratigraphic analysis of different seismic data sets and
444 the modelling of cyclic stratigraphic sequences. It is based on the identification of dated
445 erosion paleosurfaces that are interpreted as representing 100 ky glacial cycles (Fig. 5). The
446 present position of the Messinian erosion surface is consistent with this value, and this work
447 was used as a reference for estimating the vertical evolution of the margin (Fig. 6).

448

449 For the same period, Burrus and Audebert (1990) estimate the tectonic subsidence from about
450 20 m/My on the continental platform to about 180 m/My in the deep basin. The basic
451 mechanism for postrift subsidence is thermal relaxation. However, according to Bessis and
452 Burrus (1986), the loading effect of the sediment would contribute by 40 to 50 % to the total
453 subsidence of the margin. Consequently, part of the high increase in accommodation could be
454 due to the loading effect of sediments. The water loading effect was investigated by Lambeck
455 and Bard (2000) on the basis of a comparison between observational evidence for sea-level
456 changes along the French Mediterranean coast and the prediction from a glacio-hydro-
457 isostatic model. From the last glacial period to present-day (Fig. 6), they tested the impact of
458 the sea-level rise on the margin. A difference of 15 m between the position of sea-level during
459 LGM and the present-day position of this paleoshoreline, confirms the importance of the
460 isostatic rebound due to decreasing water column (Lambeck and Bard, 2000).

461

462 **4.3 Results**

463

464 ***Insert Table 1***

465

466 Based on these estimates of the geohistory subsidence and using a compilation of global sea
467 level from Waelbroeck *et al.* (2002) (Fig. 6), we ran several isostatic models with *Sedflux* for
468 the last climatic cycle from 125 ky (MIS-5) to present-day (MIS-1) (Fig. 2). Models R1-R5
469 (Table 1) represent different isostatic adjustments obtained from the range of parameters
470 tested and are broadly consistent with isostatic effective elastic thickness (*EET*) and relaxation
471 time (*RT*), described in similar studies (sect. 4.2). The *EET* allows us to set local to regional
472 flexural isostatic compensation and *RT* to fix slow to fast margin adjustments (Table 1). The

473 geohistory subsidence (*GS*), used for simulation input, was defined as a progressive seaward
474 tilt, taking into account the measured values of 255 m/My at 70 km from the coast (Rabineau,
475 2001). The convergence point of major seismic discontinuities roughly corresponds to the
476 present shoreline (50 km on the simulated section). In a first approximation, this convergence
477 point represents the position of the tectonic hinge point. However, its precise location cannot
478 be determined geometrically because the magnitude of erosion affecting each surface is
479 unknown. The sediment load (*SL*) effect is included in *GS*. Therefore, the R1-R5 Models were
480 run without sediment input. Only the sea-level variations (*WL*) were added to *GS* in order to
481 simulate the total subsidence (*S*). The simulation duration corresponds to the last 125 ky from
482 the last interglacial and high sea-level to present warm period and highstand (Fig. 6). The
483 strategy for these simulations is to compare the interglacial position of two successive
484 erosional surfaces formed during two successive glacial low sea-levels.

485

486 ***Insert Figure 6***

487

488 The seismic discontinuity *D70* (formed during MIS-2) is presently observed at the position
489 that corresponds to interglacial MIS-1. *D70* is used as the initial surface for the simulation.
490 We make the assumption that this surface represents the closest position that was occupied by
491 the previous seismic discontinuity *D60* (formed at the penultimate glacial period MIS-6)
492 during the last interglacial MIS-5 (125 ky). The final surface at the simulation end is
493 compared with the present-day position of *D60* on seismic profiles. Otherwise, as described
494 on seismic profiles, sandy wedges with high-angle clinoforms are preserved on the outer shelf
495 and represent successive glacial shorefaces that can be used as a “dipstick” for sea-levels. It
496 must be noted that the magnitude of erosion of these deposits was different during the last two
497 glacial cycles; the last glacial sandy shoreface being better preserved compared to the

498 shoreface formed during MIS-6 (Fig. 6). Nevertheless, the adjustment of the final simulated
499 surface, based on these features, is feasible.

500

501 ***4.3.1 Test of rheology***

502

503 The results of different tests on the lithospheric and asthenospheric behaviour are presented in
504 figure 7. For each isostatic parameter (*EET* and *RT*) and for different values of them, the
505 initial and final surfaces are plotted. Figure 7 shows, successively, the initial surface at the
506 high sea-level of simulation onset (125 ky), the position of this surface during glacial low sea
507 level, and finally, the simulated final surface at present-day position. The last erosional
508 surface is compared with the present-day position of *D60* observed on the seismic profile
509 (Fig. 5) and reported in this graph.

510

511 ***Insert Figure 7***

512

513 We tested *EET* ranging from 50 to 100 km. Our results show that this value has relatively
514 limited impact on the position of the final observed surface (Fig. 7). Only a difference in the
515 isostatic response during the glacial period can be observed. In contrast, large variations of the
516 position of final surfaces on the basis of different *RT* confirm the importance of this semi-
517 empirical parameter, controlled by upper mantle cooling (sect. 2.2.2). The difference between
518 the simulated and observed final position of *D60* is associated with the morphological
519 difference between *D70*, used as the initial surface, and *D60*, the modelled surface (Fig. 7).
520 The comparison is made to sandy paleoshorefaces that mark the paleoshoreline position
521 through time. There is a good match between the simulated and observed paleoshoreline,
522 except for differences in erosion. At this stage, the simulations do not take this erosion into

523 account. From these tests, the preferred isostatic adjustment R8, for the following simulations,
524 use an effective elastic thickness of 65 km and a relaxation time of 3,500 years (Table 1). The
525 seismic interpretation otherwise leads us to run different scenarios of the geohistory
526 subsidence. In particular, the seaward migration of the convergence point between *D70* and
527 *D60* from 50 to 80 km provides a best fit between simulated and observed final surfaces
528 between 70 and 90 km on the working section. Note that above this point, simulations are not
529 precisely constrained. Only the marine part of the model is accurate enough for simulating the
530 vertical motions.

531

532 ***4.3.2 Evolution of the total subsidence through time***

533

534 Stratigraphic modelling points to a specific position along the section through the 125 ky of
535 simulations. We observed, in particular, the evolution of the elevation at 100 km, where the
536 total subsidence (*S*) variations can be quantified (Fig. 8). Simulations have taken into account
537 only the geohistory subsidence (*GS*) or the water loading (*WL*) effect. Therefore, *S*, *WL* and
538 *GS* are plotted as a function of simulated time and the difference of evolutions can be
539 monitored (Fig. 8). The modelling confirms that the simple addition of *WL* and *GS* is not
540 sufficient to reproduce the total subsidence. The evolution of each component of subsidence is
541 dependant on the others. In particular, the water loading is mainly the consequence of relative
542 sea-level, which is partly dependant on the geohistory subsidence (Fig. 1). The second aspect,
543 deduced from these observations, is to consider the rapid total subsidence variations as the
544 results of the water loading fluctuations. The *GS* is assumed to be constant along the
545 simulations; the sea-level oscillations are the only parameter that can modify the load on the
546 shelf. As a result, total subsidence mimics the seal-level variations of the last climatic cycle
547 (Fig. 8.2).

548

549 **Insert Figure 8**

550

551 ***4.3.3 The geohistory subsidence and water loading effect***

552

553 We estimate the importance of water loading on the vertical evolution of the shelf. From the
554 tests described in figure 7, the platform is uplifted as sea-level falls until the glacial period
555 (21 ky) even with geohistory subsidence active. Confirmation is seen in the results from the
556 adapted isostatic model R8 (Fig. 8); the elevation, which only takes into account water
557 loading, rises between 125 and 21 ky and then rapidly drops after the Glacial period when
558 relatively low sea-level unloads the shelf. The impact of the *WL* on the shelf, between the
559 glacial sea-level lowstand (21 ky) and the present-day highstand, can cause isostatic sinking
560 of about 20 m at 100 km on the simulated section (Fig. 8).

561

562 Similarly the progressive sinking of the margin due to the combination of the tectonic/thermal
563 subsidence and the sediment loading (*GS*) is estimated to be about 15 m for the last climatic
564 cycle and about 5 m for the last deglaciation (from 21 ky to present-day).

565

566 ***4.3.4 Estimation of the sediment loading effect***

567

568 Tectonic/thermal subsidence and sediment loading both contribute to the *GS*, which has been
569 evaluated as 15 m of subsidence during the last climatic cycle (Fig. 8). During the
570 stratigraphic model simulation R8, the sediment load contribution is taken into account in the
571 *GS*, as we did not add sediment into the model domain. In order to estimate the contribution
572 of sediment load into the *GS*, the model R9 was run with a sediment source (Table 1). The

573 amount of sediment input was determined in order to obtain a sedimentary thickness
574 equivalent to that measured on the seismic interpretation.

575

576 For this study, we considered the sediment influx as a constant parameter during the 125 ky;
577 locally, the change in accommodation and the sediment load could suffer from this
578 hypothesis, but the isostatic model, used to simulate the behaviour of the margin, is a regional
579 flexural model and local loads do not have a significant impact on the isostatic response.
580 Moreover, the modelled sedimentary thickness allows us to compare and to adjust the seafloor
581 from seismic data and simulations at the top of this wedge.

582

583 Because sediment load is already included in the *GS*, model R9 calculates the final position of
584 the erosion surface with twice the sediment load; one with the *GS* estimate and one with an
585 imposed sediment flux. The final result allows us to present the simulated *D60* (with $2.SL$)
586 and the simulated *D70*, as we modelled the sedimentary thickness between these two major
587 erosion surfaces (Fig. 9). The difference between simulated *D60* from R8 (*SL* is comprised in
588 *GS*) and from R9 ($2xSL$) provides about 5 more meters of total subsidence ($S+SL$) at the end
589 of the last climatic cycle simulation (Fig. 9), compared to the previous R8 total subsidence
590 (*S*). It is mainly from the added *SL* contribution. If we now consider the 15 m of *GS* for the
591 last climatic cycle, we can infer about 10 m of sinking related to the tectonic and thermal
592 subsidence. Consequently, and from these 125 ky simulations, about 1/3 of *GS* is a result of
593 sediment load (*SL*), while the remaining 2/3 is due to thermal subsidence (*TS*).

594

595 **Insert Figure 9**

596

597 **4.4 Discussion**

598

599 Successive low sea-level during glacial periods and their corresponding erosional surfaces
600 have a cyclic repetition through time in the Gulf of Lions. Using these surfaces as chrono-
601 stratigraphic indicators, it is possible to estimate the geohistory subsidence (*GS*) and the
602 stratigraphic simulations of the last climatic cycle. In the *Sedflux* model, the flexural isostatic
603 adjustments and compensation were simulated using parameters that are broadly consistent
604 with isostatic effective elastic thickness (*EET*) and relaxation time (*RT*) described in similar
605 studies (Huybrechts, 2002; Lambeck and Bard, 2000; Peltier, 1998; Watts, 2001). The high
606 values of *EET* used for these experiments can be compared to the global bi-modal distribution
607 of continental *EET* values established by Watts (1992). The 65 km effective elastic thickness
608 corresponds to the lower part of Watts' distribution, that describes basins generally developed
609 on high *EET* cratonic interiors. This value is in agreement with the geological setting of the
610 structural region; the Gulf of Lions is a passive margin originating from Alpine thrust and its
611 lithospheric thickness results from a long and complex marine and continental geological
612 history.

613

614 **Insert Table 2**

615

616 From the simulations, we estimate each component of the total subsidence (*S*) (Table 2 and
617 Fig. 9). First, the water loading (*WL*), associated with relative sea-level variations, can impact
618 the vertical evolution of the margin between glacial lowstand and interglacial highstand. The
619 range of *WL* effect can reach about 20 m. As a consequence, the position of glacial
620 sedimentary features, observed at the present-day outer shelf, have to be corrected from this
621 subsidence in order to approach the real water depth of their formation. For instance,
622 Lambeck and Bard. (2000) determined a 15 m isostatic rebound for the last deglaciation

623 period in the Gulf of Lions, as compared to our value of 20 m. Second, the geohistory
624 subsidence (*GS*) measured on seismic is confirmed. The thermal cooling effect (*TS*) of the
625 margin would contribute to about 60 to 65 % of the *GS*; the remaining fraction being
626 represented by the sediment load (*SL*). These estimates are similar to those defined by Bessis
627 and Burrus (1986). They are in the same range as estimates for other areas found in the
628 literature (sect. 4.2) (Fig. 2). These comparisons demonstrate that our method provides, for
629 short time-periods, results in the same range as those obtained through traditional approaches
630 for longer time-scales. It allows one to take into account the hydro-isostatic effect, which is a
631 key parameter for studying the impact of Quaternary glacial-interglacial sea-level changes.
632 Moreover, our simulations confirm that the first order controls on depositional patterns are
633 sea-level change and sediment supply, but that accommodation determines what is preserved
634 in the Quaternary stratigraphic record.

635

636 **5. Conclusion**

637

638 Subsidence corresponds to the movement of the Earth's surface with respect to a reference
639 level. Total subsidence has two major components: a tectonic part (mechanic and thermal)
640 and a gravity part (sedimentary and hydrostatic loading). The aim of our method allows each
641 of these components to be evaluated from a combined field stratigraphic and modelling
642 approach. One application of the method is to estimate the impacts of global sea-level
643 changes for a site where relative sea-level changes are stratigraphically well-constrained. The
644 amplitudes of global Quaternary sea-level oscillations, derived from paleoclimatic proxies
645 obtained from ice or sediment cores, need to be calibrated by independent geological
646 measurements. On passive margins, the shelf edge is the place where accommodation is most

647 important and direct measurement of successive relative sea-level positions is possible from
648 the stratigraphic record (Jouet *et al.*, 2006; Rabineau *et al.*, 2005; Skene *et al.*, 1998).
649 The only way to evaluate absolute sea-level positions is to estimate the subsidence and correct
650 the local eustatic curve. With new estimates of geohistory subsidence (*GS*) and global sea-
651 level variations, stratigraphic simulations are adapted to understand the impact of water and
652 sediment loading (*WL* and *SL*) on the shelf. Their rapid fluctuations (100 ky cycles) are
653 superimposed on an overall trend corresponding to thermal cooling (*TS*) of the continental
654 margin. This tectonic subsidence is a portion of the vertical deflection of the Earth's surface
655 through time due to basin formation.

656

657 The stratigraphic modelling of the Rhône deltaic margin during the last climatic cycle
658 (125 ky) allows the assessment of parameters estimated with the geohistory analysis
659 (tectonic/thermal subsidence and sediment loading). Global eustasy fluctuations using the
660 *Sedflux* model provides confirmation of the important impact of water loading on vertical
661 motions of the platform between glacial low sea level and interglacial high sea level. Finally,
662 the thermal subsidence and the sediment load both contribute to the geohistory subsidence,
663 defined from stratigraphic analysis, with a relative impact of 60-65 % and 35-40 %,
664 respectively.

665

666 From seismic profiles and their interpretation, stratigraphic modelling provides a way to
667 either confirm or discard different hypothesis on subsidence. This method permits us to test
668 hypothesis without some of the information needed for the "backstripping" method; for
669 example lithological information is unnecessary. However, several assumptions, considered
670 as reasonable in the case of our study area, have to be made, especially about the repetition of
671 the sedimentary features (glacial erosion surfaces) through each glacial cycle. A future

672 application of this technique will be to better constrain sediment supply. The European project
673 PROMESS, which cored two boreholes on the outer shelf in 2004, will provide lithological
674 and geochronological information for the last 500 ky sedimentary record. A new analysis at
675 this time-scale will soon be undergone, taking into account the entire Gulf of Lions platform,
676 in order to assess the space and time variability of the different components of subsidence.

677

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679

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690

691 **7. References**

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853 **8. Figure captions**

854

855 **Fig. 1:** Definition of mechanisms causing subsidence on a passive continental margin. Total
856 subsidence (*S*) results from vertical motions that are specific for each lithology. Different
857 subsidence components can be determined by mechanisms that cause these motions. Note
858 term geohistory subsidence (*GS*) that corresponds to subsidence measured from seismic
859 stratigraphy and associated to combination of tectonic/thermal subsidence (*TS*) and sediment
860 load (*SL*).

861

862 **Fig. 2:** Flow chart of method defined in this paper and aimed at investigating different
863 components of total subsidence (*S*). Strategy uses stratigraphic modelling along with *Sedflux*
864 program and input parameters from geohistory and stratigraphic analysis.

865

866 **Fig. 3:** Gulf of Lions continental margin (North-Western Mediterranean); Geographic,
867 morpho-bathymetric (from Berné *et al.*, 2002) and hydrographic settings. Dotted line
868 indicates position of NE–SW seismic synthesis shown on figure 4. Location of North-South
869 simulated profile on map shows that marine section presented here is between 50 and 130 km.
870 Note that simulations take into account entire profile (including onshore section).

871

872 **Fig. 4:** Stratigraphic interpretation from composite high-resolution Sparker seismic lines
873 (position in Fig. 3). NE–SW transect across platform illustrates stacking of last five
874 sedimentary sequences (S1 to S5) bounded by major discontinuities (D40 to D70). Within
875 these sequences, deposits are organized in a vertically stacked sedimentary motif consisting of
876 prisms (PI) with gently dipping clinoforms, and prisms (PII) with relatively high-angle

877 clinofolds (from 3 to 7°) (Rabineau *et al*, 2005). Highest amplitude seismic reflections reveal
878 major erosional surfaces that formed during overall sea-level fall and lowstands.

879

880 **Fig. 5:** Chirp seismic profile across continental shelf (position in Fig. 3). Major seismic
881 surfaces correspond to cyclic erosion surfaces formed during forced regressions (see text for
882 detailed explanation).

883

884 **Fig. 6:** Definition of geohistory subsidence (*GS*) and water loading (*WL*) as they are input in
885 numerical stratigraphic model *Sedflux*. Identification of dated erosion paleosurfaces permits
886 one to quantify their vertical evolution through time, and to estimate *GS* subsidence. This
887 value takes into account both tectonic subsidence (*TS*) and loads due to sediment deposition
888 (*SL*). Water loading (*WL*) results from relative sea-level fluctuations and reaches its maximum
889 between glacial and interglacial periods.

890

891 **Fig. 7:** Test of isostatic parameters used in numerical stratigraphic model *Sedflux*. Total
892 subsidence (*S*) is modelled using different values of **1-** effective elastic thickness (EET) and
893 **2-** relaxation time (RT).

894

895 **Fig. 8: 1-** Modelling of total subsidence (*S*) with best adapted isostatic parameters (Model R8:
896 EET = 65 km; RT = 3,500 years). **2-** Evolution through time of *GS*, *S* and *WL* subsidence
897 from this simulation R8.

898

899 **Fig. 9:** Modelling of total subsidence (*S*) and (*S'*). (*S'*) is results of Model R9 with sediment
900 flux turned on. Comparison of these models permits estimation of different components of
901 total subsidence (*S*) at 100 km of simulated section (outer shelf).

902

903 **Table 1:** Description of different numerical stratigraphic models (R1 to R9) and their input
904 parameters.

905

906 **Table 2:** Quantification of different components of total subsidence (S) at 100 and 120 km of
907 simulated section.

908

909

PASSIVE CONTINENTAL MARGIN

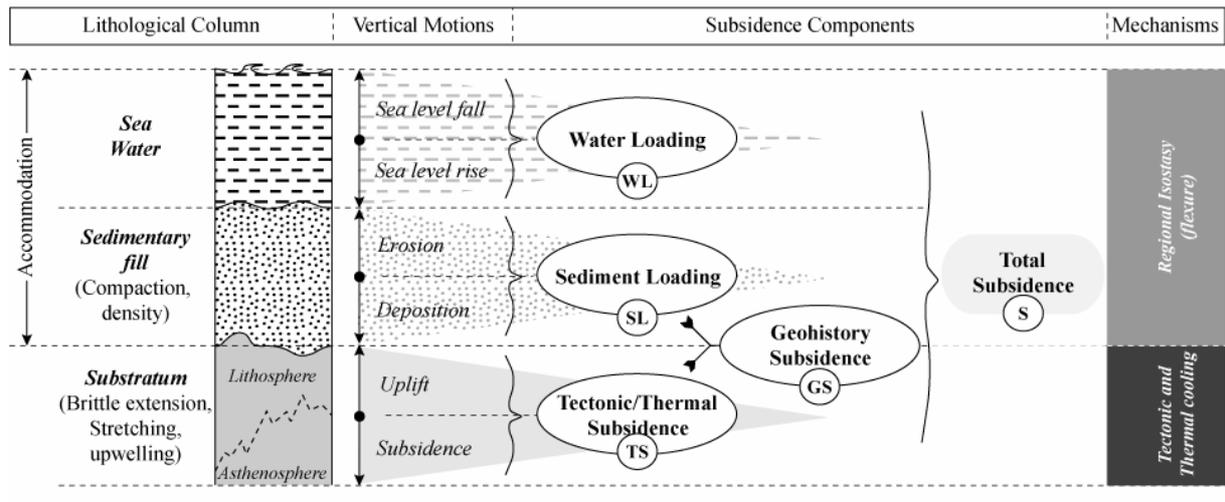


Figure. 1 - Jouet *et al.*

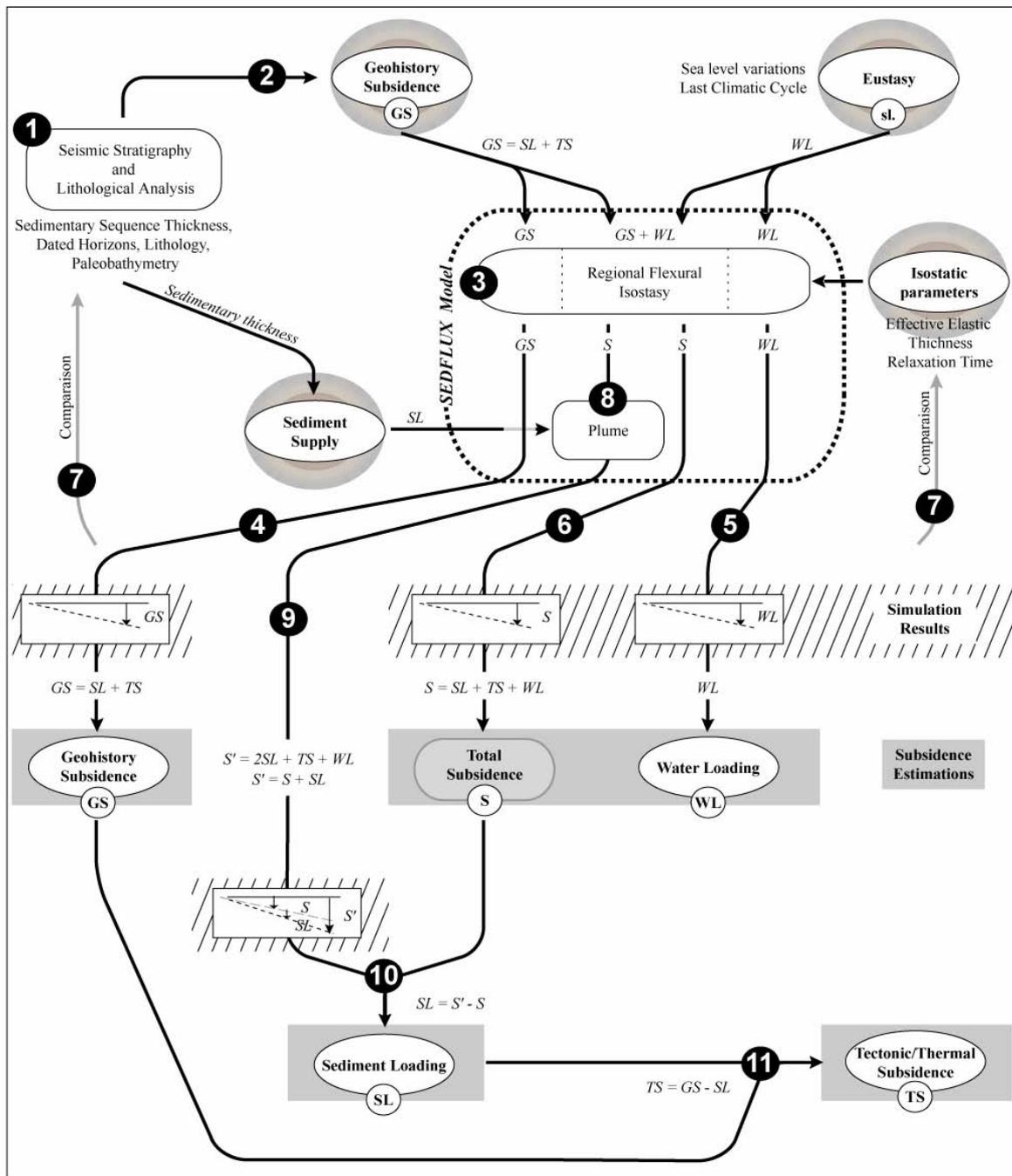


Figure. 2 - Jouet et al.

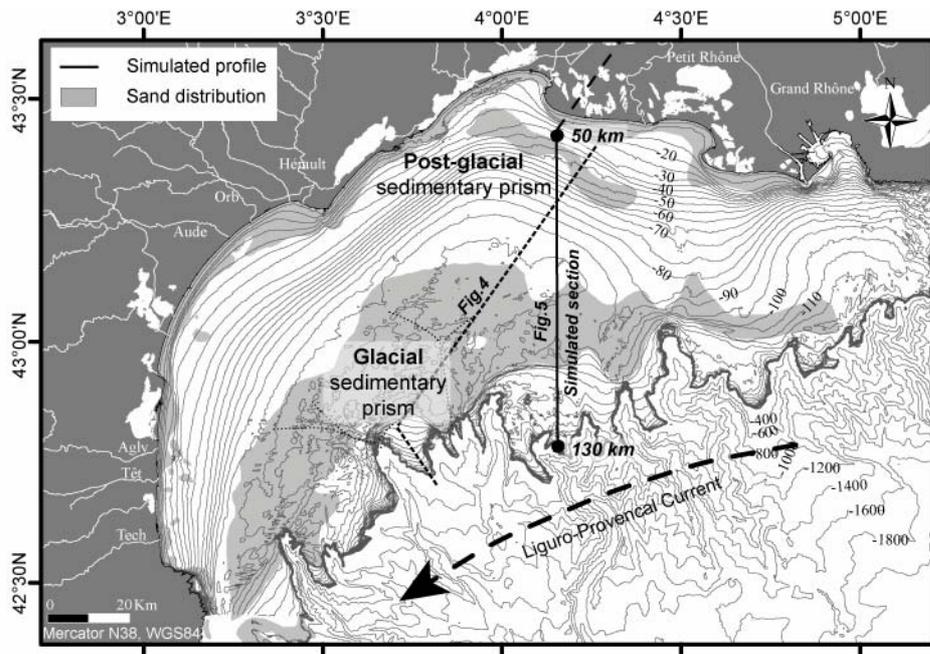


Figure. 3 - Jouet *et al.*

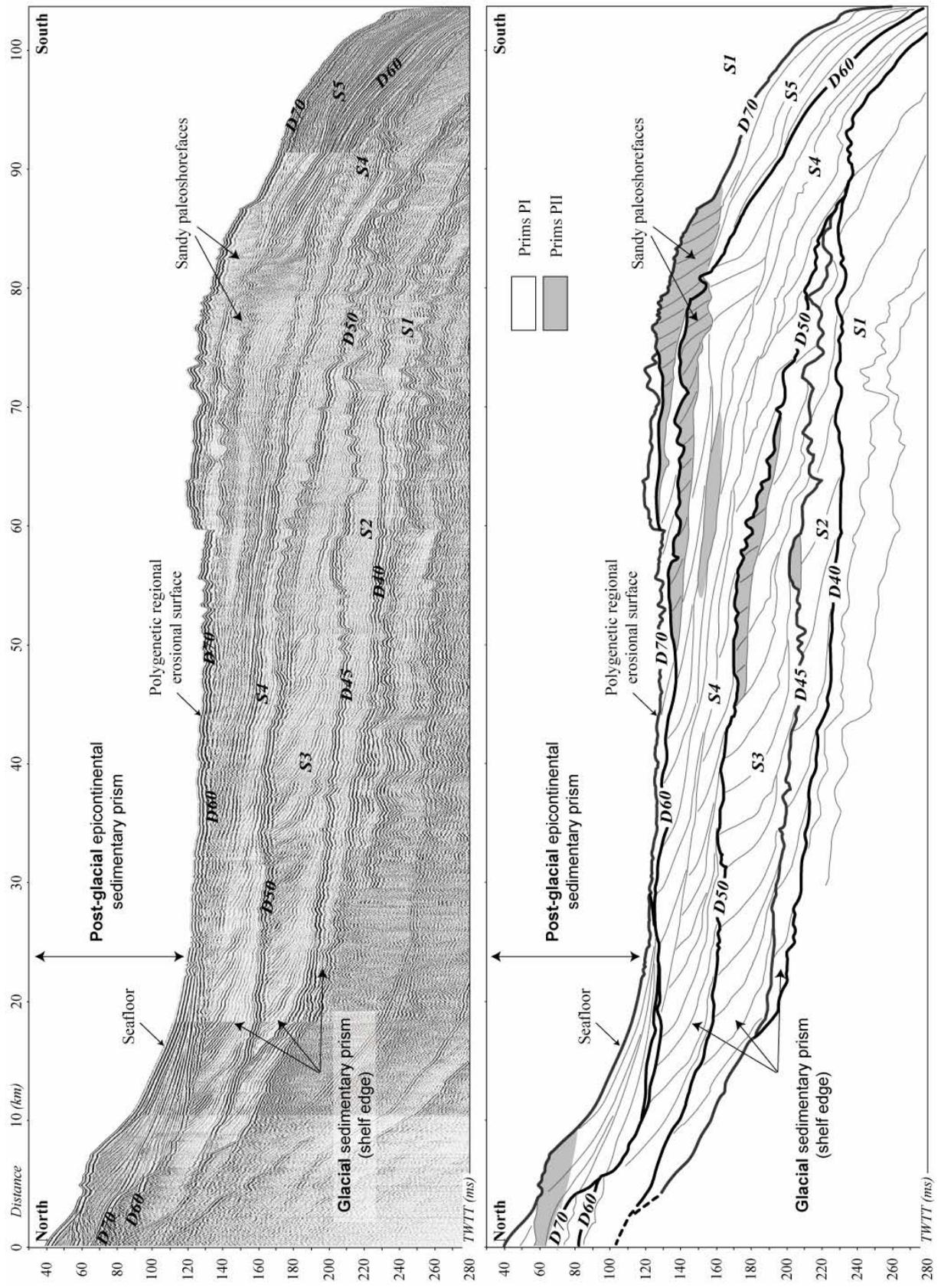


Figure. 4 - Jouet *et al.*

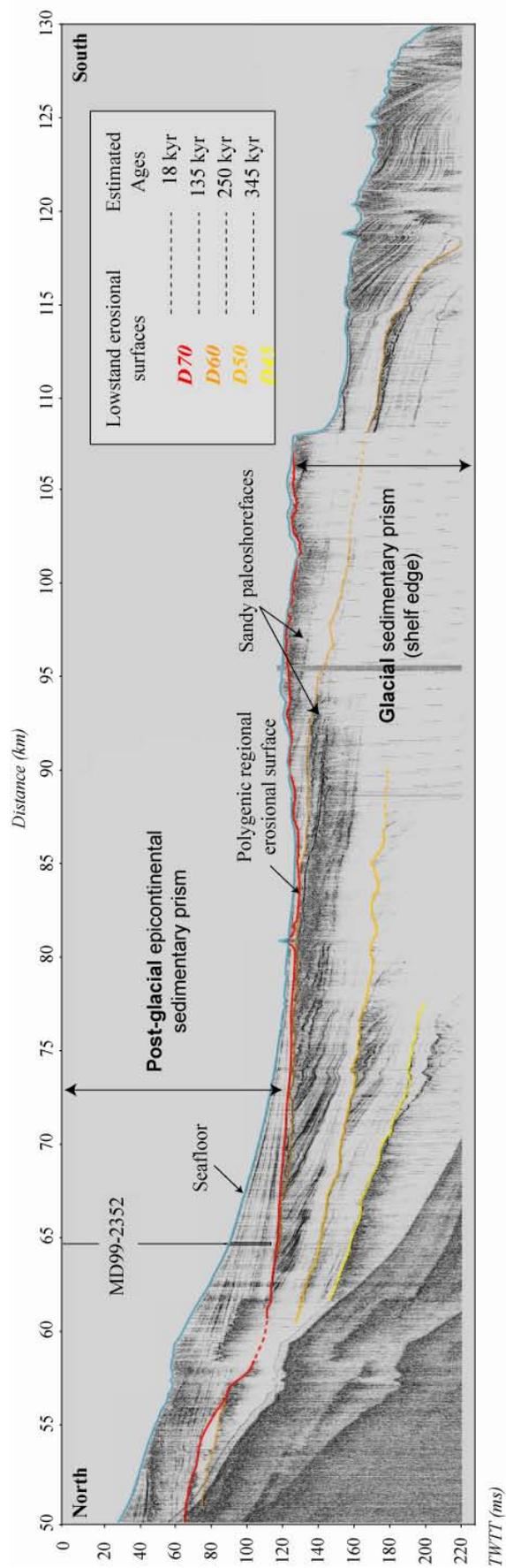


Figure. 5 - Jouet *et al.*

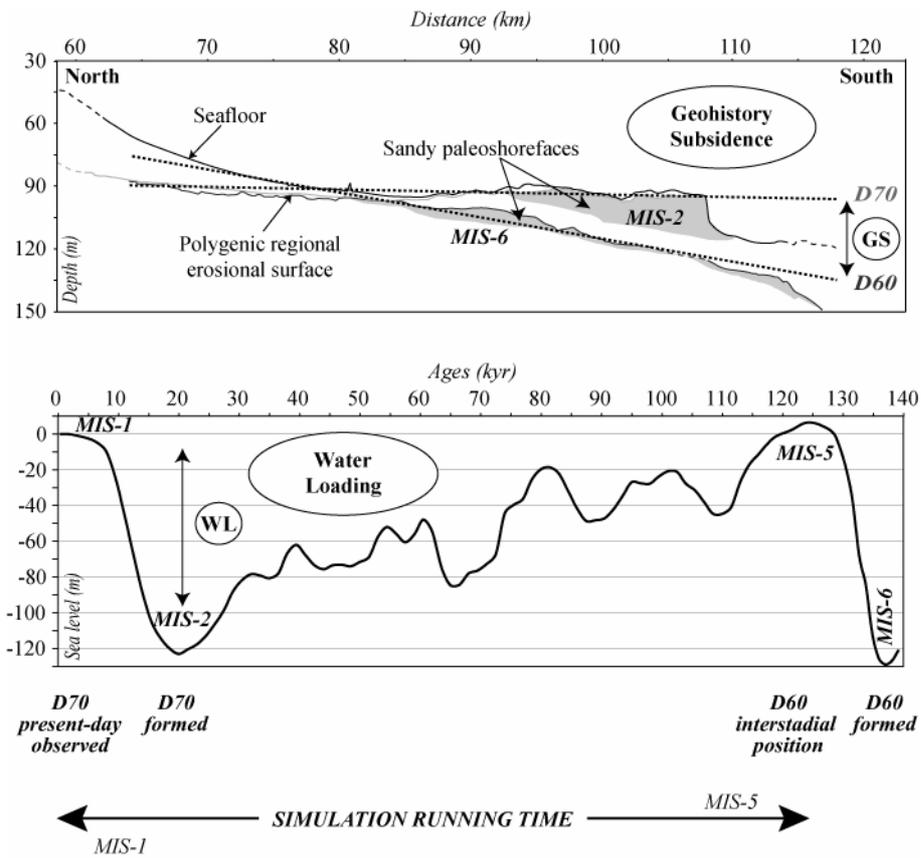


Figure. 6 - Jouet *et al.*

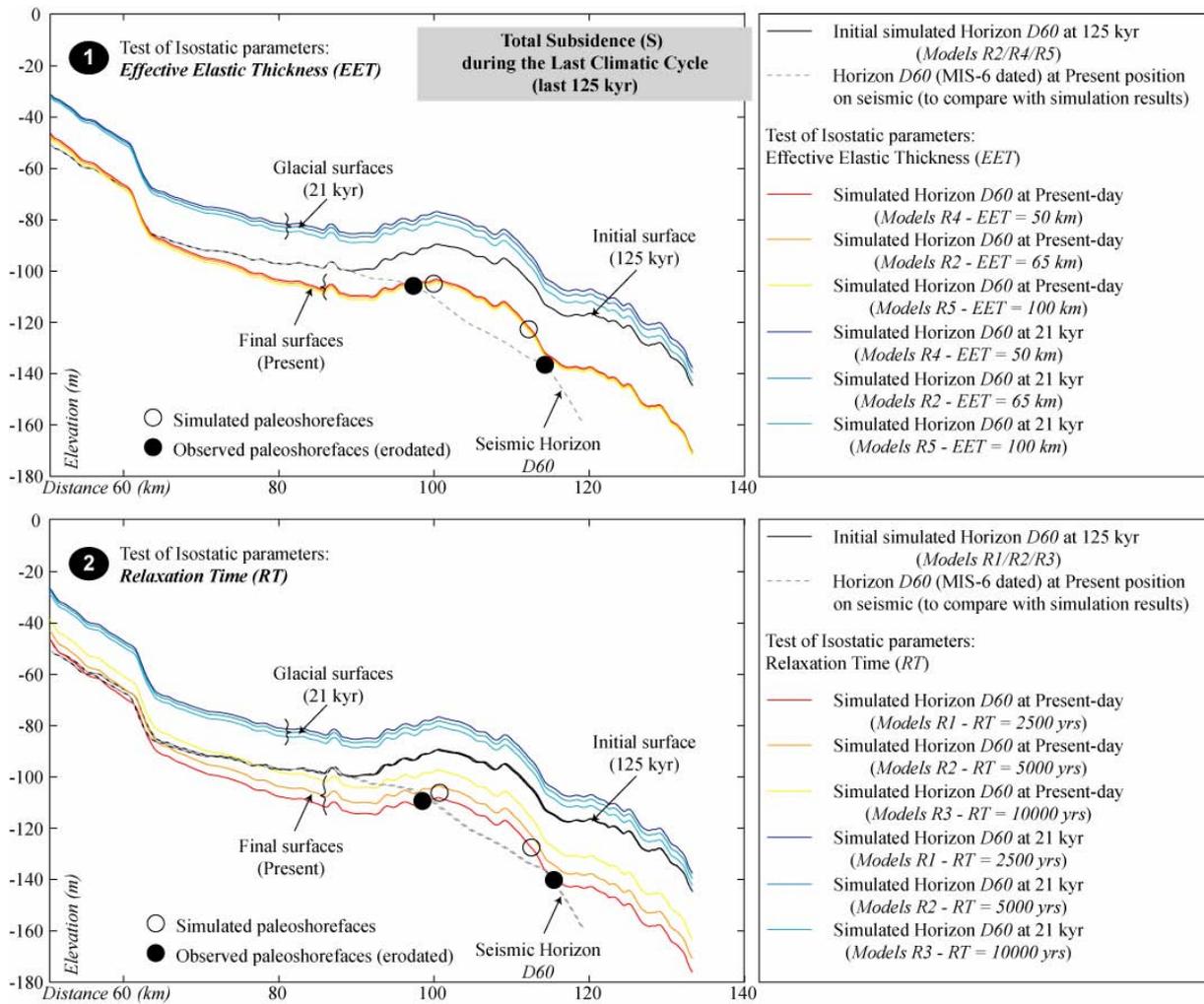


Figure. 7 - Jouet *et al.*

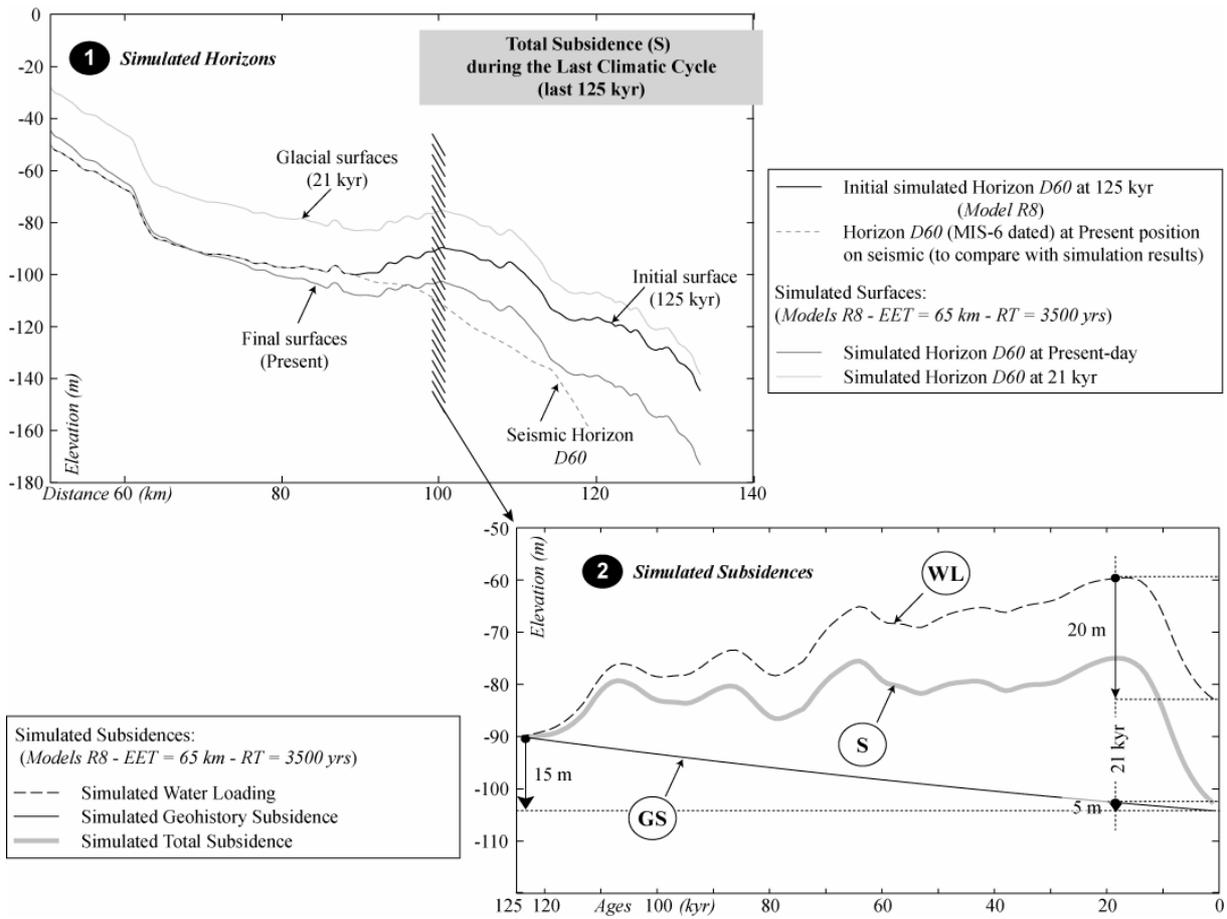


Figure. 8 - Jouet et al.

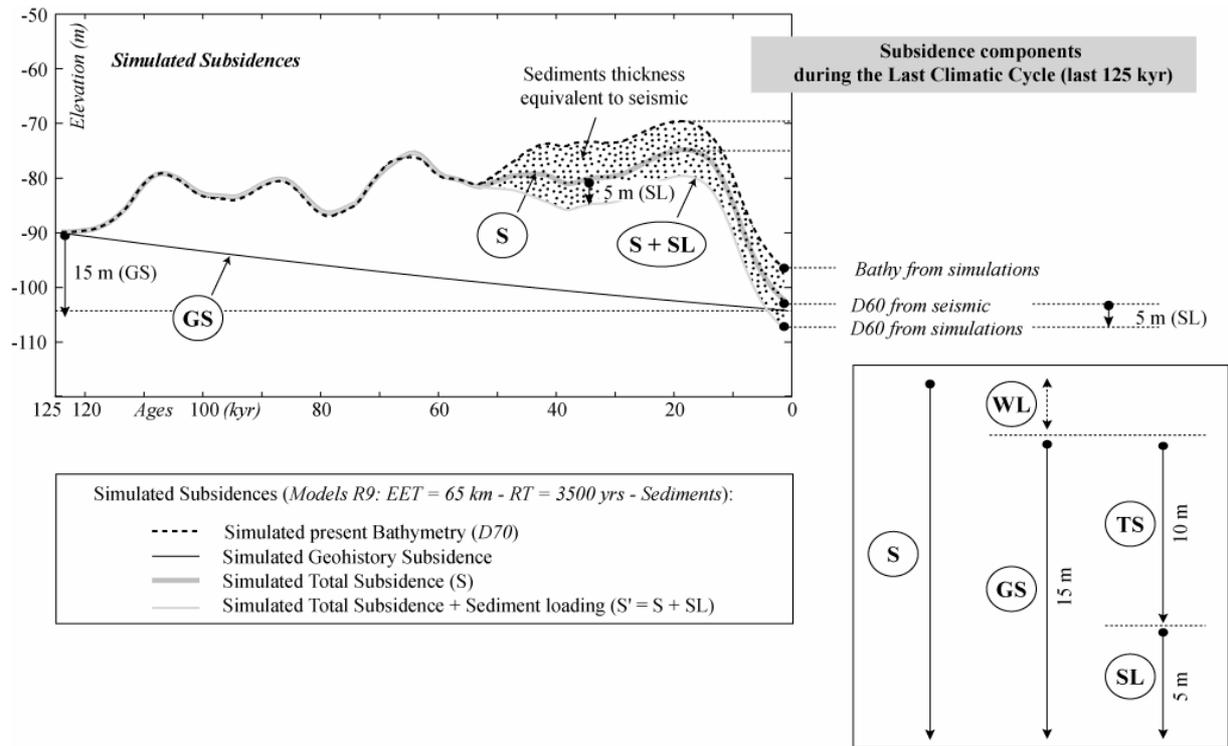


Figure. 9 - Jouet et al.

MODELS	DESCRIPTION	ISOSTASY		SUBSIDENCE		SEDIMENT	DURATION
		<i>Isostatic Effective Elastic Thickness (km)</i>	<i>Isostatic Relaxation Time (yrs)</i>	<i>Inner shelf (m/yr)</i>	<i>Outer shelf (m/yr)</i>		
R1	Fast margin adjustment	65	2 500	0 (50 km)	2,5E-04	No	LCC (125 kyr)
R2	Medium margin adjustment	65	5 000	0 (50 km)	2,5E-04	No	LCC (125 kyr)
R3	Slow margin adjustment	65	10 000	0 (50 km)	2,5E-04	No	LCC (125 kyr)
R4	Local margin adjustment	50	5 000	0 (50 km)	2,5E-04	No	LCC (125 kyr)
R5	Regional margin adjustment	100	5 000	0 (50 km)	2,5E-04	No	LCC (125 kyr)
R6	Conv. point migration, fast adjustment	65	2 500	0 (50-80 km)	2,5E-04	No	LCC (125 kyr)
R7	Conv. point migration, medium adjustment	65	5 000	0 (50-80 km)	2,5E-04	No	LCC (125 kyr)
R8	Adapted eath model (LCC)	65	3 500	0 (50-80 km)	2,5E-04	No	LCC (125 kyr)
R9	Estimation of the Sediment load	65	3 500	0 (50-80 km)	2,5E-04	Yes	LCC (125 kyr)

Table. 1 - Jouet *et al.*

MODELS	DESCRIPTION	POSITION 100 km						
		Total Subsidence LCC D60 (m)	Sediment Thickness D60-D70 (m)	Total Subsidence LD D70 (m)	Subsidence (Sed. Load. + Teconics) (m)	Hydro-Isostasy (Water Load.) (m)	Sediment Loading (m)	Tectonic Subsidence (m)
- At position 100 km -								
R8	Adapted eath model (LCC)	+15	+10	+25	+4,5	+20,5		
R9	Estimation of the Sediment load						1/3 Subsid	2/3 Subsid
- At position 120 km -								
R8	Adapted eath model (LCC)	+10	+15	+25	+5,5	+19,5		
R9	Estimation of the Sediment load						1/3 Subsid	2/3 Subsid

Table. 2 - Jouet *et al.*