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Sedimentary and structural record of the Albian growth of the Bakio salt diapir (the Basque Country, northern Spain)

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Abstract

Whereas salt has a viscous rheology, overburden rocks adjacent to salt diapirs have a brittle rheology. Evidence of deformation within the overburden has been described from diapirs worldwide. Gravity-driven deposits are also present along the flanks of several diapirs. The well-known example from the La Popa Basin in northern Mexico shows that such deposits may be organized into halokinetic sequences.

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This leads to the several questions: (i) How does diapir growth contribute to overburden deformation? (ii) Are halokinetic sequence models valid for other areas beyond the La Popa Basin. The Bakio diapir and its well-exposed overburden in Basque Country, Spain, provides key elements to address these questions.

The Bakio diapir consists of Triassic red clays and gypsum and is flanked by synkinematic middle to upper Albian units that thin toward the diapir. The elongate diapir parallels the Gaztelugatxe normal fault to the NE: both strike NE-SW and probably formed together during the middle Albian, as synkinematic units onlap the fault scarp. The diapir is interpreted as a reactive diapir in response to middle Albian motion on the Gaztelugatxe fault. The rate of salt rise is estimated to be about 500 m/Myr during this passive stage. During Late Albian, the diapir evolved passively as the Gaztelugatxe fault became inactive. Synkinematic units thinning toward the diapir, major unconformities, slumps and other gravity-driven deposits demonstrate that most deformation related to diapir growth occurred at the sea floor. Halokinetic sequences composed of alternating breccias and fine-grained turbidites recorded cyclic episodes of diapir flank destabilization.

This work provides insights into drape fold and halokinetic sequence models and offers a new simple method for estimating rates of diapir growth. This method may be useful for outcrop studies where biostratigraphic data are available and for other passive diapirs worldwide.

Keywords : diapir, halokinetic sequences, drape folds, drag folds, slumps, gravity-driven deposits.

1 Introduction

Steep, vertical or even overturned strata associated with thinning of strata toward growing salt features are commonly described adjacent to diapir walls (e.g. Jackson et al. 1994; Alsop et al. 2000; Giles and Lawton, 2002; Rowan et al., 2003; Schultz-Ela, 2003; Banham and Mountney, 2013a, b, c). Such structures are related to overburden deformation during diapir growth and are commonly difficult to study. In the field, late dissolution of evaporite masses (commonly gypsum or halite) often triggers collapse of the adjacent overburden (e.g. Jackson et al., 1998). On seismic lines, resolution is generally poor near diapir walls because of steep dipping beds, faulted strata, overhanging salt, extreme stratigraphic thickness changes and poor velocity control (Davison et al., 2000). Two different models have been proposed to explain these structures found along diapir flanks (Fig. 1): drag fold (Jackson and Talbot, 1991; Jackson et al., 1994) and drape fold models (Schultz-Ela, 2003). Drag folds involve shearing, folding and faulting of the overburden by evaporite rise with deformation at depth. Drape folds correspond to folding at the basin floor due to differential vertical motion between the diapir and its overburden. Vertical motions of the basin floor may create sedimentary sequences controlled by the

balance between the rates of diapir rise and adjacent sedimentation. Such halokinetic sequences have been described in the El Papalote diapir (La Popa basin, northern Mexico; Giles and Lawton, 2002 ; Rowan et al., 2003).

This paper presents field-based sedimentological and structural analyses from the Bakio diapir in the Basque-Cantabrian Basin. The Bakio overburden displays syn-diapir growth strata, well-exposed sedimentary wedges that thin toward the diapir, major unconformities and upturned strata. The aim here is to describe and discuss the deformation patterns and sedimentary response of strata related to the diapir growth. This aim will be achieved by addressing: (i) how and when did the Bakio diapir rise? (ii) how did diapir growth induce overburden deformation? (iii) Are halokinetic sequence models restricted to the El Papalote diapir or relevant for other diapirs worldwide?

The Bakio diapir provides important outcrop-based data, which is relatively rare in the literature and represents a good analogue for petroleum exploration of diapir flanks where poor seismic imaging makes interpretation difficult. This study improves understanding of the timing of the Bakio diapir growth, and provides insights for the drape fold model. This study also shows the interaction of regional and gravity-driven sedimentation triggered by the diapir within halokinetic sequences, and supports a new method for quantifying the rates of diapir growth.

2 Models for diapir formation

This section summarizes models for diapir initiation and growth, and presents the two main models of overburden deformation (drag fold versus drape) and of halokinetic sequences deposition. The term “salt diapir” is widely used in the literature to describe all salt structures with discordant contacts with its overburden, regardless of the processes of initiation and growth. The formation of salt structures may be driven by different processes such as extension, differential loading or compression. This work focuses only on diapirs developed within extensional basins.

2.1 Diapir initiation and growth

Jackson et al. (1994) demonstrated that salt diapirs display discordant contacts with the overburden, showing that diapirs pierce the overburden. As the overburden behaves more as a brittle than viscous medium (Vendeville and Jackson, 1992; Weijermars et al., 1993), Jackson et al. (1994) pointed out the space problem in which salt is able to pierce into a brittle overburden. This problem is solved by three piercement modes: active, passive and reactive diapirism (Fig. 1a). In all cases, the driving mechanism for salt flow, and consequently diapir rise, is the pressure difference between the top of a diapir and its surroundings.

Active diapir growth or piercement (Nelson, 1991) corresponds to intrusion of salt through the overburden (Jackson and Talbot, 1991) and is thought to trigger doming and faulting of the diapir roof (Fig. 1a). Viscous salt may pierce the roof if it is thin enough (Jackson et al., 1994).

Passive diapirism (Jackson et al., 1994), also called downbuilding (Barton, 1933), occurs when salt emerges at surface (Fig. 1a). Deposits accumulate around the diapir and wedge out above the diapir. In this model, the diapir shape is controlled by the balance between the rates of sediment accumulation and diapir rise (Vendeville and Jackson, 1993). When the rates are equal, the diapir grows vertically. When the diapir rises faster than the surrounding sediments deposit, salt may reach the basin floor and spread laterally. By contrast, when sediments aggrade faster than salt rise, sediments onlap and progressively cover the outcropping salt; in this case, the diapir may cease rising because of the sediment load above the diapir.

Reactive diapirism corresponds to diapir salt rising in response to regional tectonics (Fig. 1a). Jackson et al. (1994) and Vendeville and Jackson (1992) demonstrated that reactive diapirs develop below extensional half grabens or grabens (Fig. 1a). The balance between faulting and sedimentation may controll diapir initiation, as normal faults thin the overburden yet sedimentation increases the thickness of the overburden. Rapid extension with low sedimentation rates leads to progressive thinning of the diapir roof and can evolve toward active and then passive diapir growth. Low extension and high sedimentation rates lead to diapir burial and normal fault growth (Vendeville and Jackson, 1993).

As viscous evaporites cannot pierce through a thick overburden, there are two main ways to initiate diapir growth: either diapirs initiate early when the overburden is thin and unconsolidated or tectonics trigger latter diapir growth by thinning the overburden (Jackson et al., 1994). Hudec et al. (2009) and Ings and Beaumont (2010) showed that early compression of a thin overburden may initiate diapir growth. In addition, Jackson et al. (1994) demonstrated that regional extension is the most effective process to trigger reactive diapirism when there is a thick overburden. This implies that many diapirs initiated first in response to compression or extension, but subsequently experienced a rapid stage of active diapirism, which permitted piercement of the roof. This allows the diapir to evolve passively. If a diapir is buried by a thick overburden (in the case of a high sedimentation rate), it may prevent further diapir growth (Jackson et al., 1994).

2.2 Diapir related deformation

Overburden deformation along diapir flanks has been widely reported in several studies (e.g. Jackson and Talbot, 1991, Jackson et al. 1994; Alsop, 1996; Alsop et al., 1995 and 2000; Giles and Lawton, 2002; Rowan et al., 2003). Steep, vertical or overturned strata are widespread, associated with rapid thinning toward the diapir (Alsop et al. 2000). The dip of adjacent strata decrease some distance away

from the salt (about 300 to 2000 m), depending on the diapir size (Schultz-Ela, 2003). Concentric and/or radial faults also affect diapir flanks (Stewart, 2006).

Two main mechanisms have been proposed to explain overburden deformation induced by salt rise: drag and drape folding (Fig. 1b, c). Drag folds or external shear zones (Jackson and Talbot, 1991; Jackson et al., 1994; Alsop, 1996; Alsop et al., 1995 and 2000) involve shearing and deformation of consolidated sediments, along the diapir edges (Fig. 1b). In contrast, drape folds (Schultz-Ela, 2003) are related to deformation of unconsolidated sediments, where folding is explained by differential vertical motion on the diapir edges, between uplift of the diapir roof and subsidence of diapir flanks (Fig. 1c). As the older stata undergo greater tilting than younger strata, this explains various degrees of tilting, unconformities and thinning of deposits toward the diapir. It also explains why slumps may develop on diapir flanks (e.g. Giles and Lawton, 2002; Rowan et al. 2003), as local steepening by drape folding may trigger slumps and other gravity-driven deposits.

2.3 Halokinetic sequences

Sedimentary packages bounded by unconformities, with lower gravity-driven deposits and upper fine-grained deposits (Fig. 1c) are described in the overburden of the El Papalote diapir (Giles and Lawton, 2002; Rowan et al. 2003). Giles and Lawton (2002) and Rowan et al. (2003) interpreted such stacked packages as the result of cyclic processes. These authors invoked the concept of halokinetic sequences controlled by drape folding. Such sequences are dictated by the balance between the rates of diapir rise and adjacent sedimentation. When diapir rise exceeds sedimentation, the diapir produces positive surface relief and upturned strata. Uplift of the diapir roof may lead to sediment sliding and redeposition; thus gravity-driven deposits may develop unconformably on eroded upturned strata. If uplift and roof destabilization are sufficient, the diapir may breach the basin floor. When sedimentation exceeds the diapir rise, the diapir may be buried. The slope of the flanks is assumed to become gentler due to sediment onlap on the dipping slope. This explains the progressive decrease of gravity-driven deposits in the upper part of the halokinetic sequence. Cycles of salt doming and onlap lead to deposition of stacked halokinetic sequences bounded by angular unconformities. Angular unconformities develop during doming associated with erosion of upturned strata. Halokinetic sequences represent cycles of passive diapirism and small-scale active diapirism when salt periodically rises and pierces the diapir roof (Rowan et al., 2003).

Giles and Rowan (2012) provided new concepts for halokinetic sequences with two end-member types: the hook and the wedge halokinetic sequences. The hook halokinetic sequences have high-angle unconformities and common gravity-driven deposits with rapid facies variations associated with narrow zones of drape folding. Hook halokinetic sequences are found when diapir rise rate exceeds sedimentation rate. The wedge halokinetic sequences are characterized by low-angle unconformities, less common gravity-driven deposits, gradational facies variations and are associated with broad zone of

drape folding. Wedge halokinetic sequences are found where sedimentation rate exceeds diapir rise rate.

3 Geologic setting

3.1 Mesozoic history of the Basque-Cantabrian Basin

The Bakio diapir belongs to the northern edge of the Basque Trough, in the northern part of the Basque-Cantabrian Basin (Fig. 2), located between the Iberian and European plates. Extension and subsidence in the basin are driven by the opening of the Bay of Biscay and rotation of Iberia during the Cretaceous (García-Mondéjar et al., 1996). The subsidence is governed by major NW-SE striking faults and minor SW-NE, N-S and E-W striking faults forming a complex pattern with several depocentres (Vicente-Bravo and Robles, 1995; García-Mondéjar et al., 1996) (Fig. 2).

The evolution of this basin may be summarized into three sedimentary stages. The first stage is characterized by limited subsidence and by thin shallow-water siliciclastic and calcareous sediments deposition, from Triassic to Barremian (Rosales et al., 2002). Triassic reds clays and gypsum accumulated during this first stage. The second stage corresponds to Aptian to Middle Albian shallow-water Urgonian limestones and marls deposition in deeper environments (Martín-Chivelet et al., 2002; García-Mondéjar et al., 2004). The last stage is characterized by an important subsidence and by the development of Late Albian to Cenomanian siliciclastic turbidites alternating with lutites. These facies are regionally called the Black Flysch units. Aptian-Albian units are locally 7000 m thick, in the main depocentre, called the Basque Trough. At present, the Basque-Cantabrian Basin corresponds to the westernmost part of the Pyrenean realm (Fig. 2), with limited N-S shortening estimated between 25 km (Gómez et al., 2002) and 40 to 50 km (Ábalos et al., 2008).

The oldest Mesozoic deposits of the Basque-Cantabrian Basin, composed of Triassic gypsum and clays crop out in several diapirs. Among these diapirs, the Estella-Lizarra, Salinas de Oro, Gulina, and Mena diapirs (García-Mondéjar et al., 1996), Pondra and Laredo Bay diapirs (López-Horgue et al., 2010) and Gernika diapir (García-Mondéjar and Robador, 1986-1987) were rising during Albian times. Several of these diapirs are located at the intersection between major Hercynian basement faults (García-Mondéjar et al., 1996) (Fig. 2), reactivated during the opening of the Basque-Cantabrian Basin. Basement faulting controlled thin-skinned tectonics and formation of halokinetic structures during Middle-Late Albian times (Bodego and Agirrezabala, in press). Early Aptian deposits thinning toward the Gernika diapir, located few kilometres eastward from the Bakio diapir, recorded diapir growth during this time (Agirrezabala and García-Mondéjar, 1989). An isolated carbonate platform developed top of this diapir, associated with slope apron facies deposited at its edge, during Early to Middle Albian times (García-Mondéjar and Robador, 1986-1987). Angular unconformities and lateral facies changes in the Bakio

overburden (García-Mondéjar and Robador, 1986-1987; Robles et al., 1988) suggest a similar evolution for both the Bakio and Gernika diapirs. The Bakio diapir is elongated along an NE-SW axis (Fig. 3a); its northernmost part is currently located offshore. The diapir surface trace coincides with the topographic depression where the village of Bakio is located, whereas surrounding relief corresponds to the overburden (Fig. 3b). The diapir is cored by Triassic red clays, gypsum and ophites enclaves (Triassic tholeiitic magmatic rocks), cropping out east of the Bakio beach.

3.2 Detailed stratigraphy around the Bakio diapir

Sedimentary rocks of the Bakio overburden include seven stratigraphic units (Fig. 4). In stratigraphic order, the Bakio marls and the Gaztelugatxe units, and the Bakio breccias Fm. correspond to the latest Urgonian deposits of the area. The Sollube, Cabo Matxitxako, Punta de Bakio and Jata units correspond to the first units of the Black Flysch Group.

The Bakio marls unit crops out only in the eastern flank of the Bakio diapir (Fig. 3a). This unit is composed of marls interbedded with thin-bedded packstones, deposited in an outer shelf setting. The Gaztelugatxe unit is found NE of the Bakio diapir, on the Gaztelugatxe and Aketxe islands, and on the mainland coast in front of Aketxe island. García-Mondéjar and Robador (1986-1987) and Robles et al. (1988) interpreted this unit as a massive carbonate platform, although several sets of fractures give a brecciated appearance to the limestone and render identification of stratigraphic surfaces difficult. The Bakio breccias Fm. is 550 m thick and is composed of breccias with floatstone blocks, either orthobreccias or parabreccias (with marly matrix), interbedded with marls and thin-bedded grainstones. Breccias, interpreted as slope apron facies, are associated with olistolithes, up to 1 m wide. The eastern flank of the diapir displays a thick accumulation of orthobreccias (about 350 m) in the lower part of the Bakio breccias Fm., and alternating parabreccias and marls in the upper part (200 m thick, Fig. 4). The western flank shows parabreccias and thin-bedded orthobreccias interbedded with marls that correspond to a lateral equivalent of the upper part in the eastern flank (Fig. 4). For García-Mondéjar and Robador (1986-1987) and Robles et al. (1988), the Gaztelugatxe unit, representing the southern part of a carbonate platform located farther north, currently offshore, is the source for breccias in the eastern flank of the diapir.

The Sollube unit mainly consists in thin-bedded, fine-grained siliciclastic turbidites interbedded with marls. The Cabo Matxitxako unit is composed of amalgamated thick-bedded coarse-grained siliciclastic turbidites. The Sollube and Cabo Matxitxako units are found in the eastern flank of the Bakio diapir (Fig. 3a). West of the diapir, the Punta de Bakio unit consists of marls interbedded with thin-bedded, fine-grained siliciclastic turbidites, beds of parabreccias with limestone blocks, quartz pebbles and reworked turbidites. The Jata unit is composed of marls alternating with fine-grained to coarse-grained siliciclastic turbidites and include slumps and breccias. The Punta de Bakio and Jata units are found only in the western flank of the Bakio diapir (Fig. 4). The Landes Massif, currently located offshore, is considered as

the source for the siliciclastic units (Voort, 1963; Rat, 1988; Robles et al., 1988; García-Mondéjar et al., 1996; Agirrezabala, 1996; Martín-Chivelet et al., 2002).

Sparse biostratigraphic data and important lateral facies variations in the Bakio area make correlations between units from either flanks of the diapir difficult. The available biostratigraphic data (Wiedmann and Boess, 1984; García-Mondéjar and Robador, 1986-1987; Robles et al. 1988; López-Horgue et al.; 2009) are summarized in figure 4. According to these data, the Urgonian units are Early to Middle Albian in age. The Bakio breccias Fm. is early Middle Albian in age (*dentatus* Zone). The Bakio marls and the Gaztelugatxe units are not directly dated. The Bakio marls unit is probably Early Albian in age, as the Bakio breccias Fm. overlies it. The Gaztelugatxe unit may be Early Albian to early Middle Albian in age as this unit is assumed to be the source for the Bakio breccias Fm.

The units of the Black Flysch Group are Middle to Late Albian in age. The Punta de Bakio unit and the Sollube unit are early Late Albian and Late Albian in age, respectively. The Jata unit (Middle to Late Albian) has been dated as early Middle Albian (*dentatus* Zone) by López-Horgue et al. (2009) further west, whereas it overlies the Punta de Bakio unit (early Late Albian) in the western flank of the Bakio diapir. The Sollube unit is probably a lateral equivalent of the Punta de Bakio and Jata units. The Cabo Matxitxako unit is not dated and may be Late Albian or Early Cenomanian in age.

The latest Urgonian unit (Bakio breccias Fm.) is early Middle Albian in age whereas overlying units of the Black Flysch Group are Late Albian in age. This defines a hiatus of most of the Middle Albian, as described by García-Mondéjar et al. (2004) and López-Horgue et al. (2009).

4 Methods

This work used field data as bedding planes, slumps axes, fault orientations measurement, stratigraphic sections building and facies analyses. Some of these methods need special explanations that are presented below.

4.1 Restoration of sedimentary structures

In this study, sea floor dip direction were inferred using several sedimentary structures. Sole marks in turbidites provided directions of turbidity flows and slumping directions may be deduced as they may be roughly perpendicular to the mean trend of slump fold axes (e.g. Alsop and Marco, 2011). Because diapir growth tilted most of beds in the overburden, a restoration was necessary. All beds with sedimentary structures have been tilted back to a horizontal position because gravitational processes occur on low-angle slopes, as little as few degrees (not higher than 10° in submarine environment). This assumption of initially horizontal bed is suitable for turbidites as such deposits develop on low angle slopes (< 1°). Slumps may deposit on steeper slopes, not higher than 10°. However, an error of about 10 degrees in the

dip of the initial slope may not modify the mean trend of slump folds axes significantly, thus may not change the associated slumping direction. Therefore the assumption of initially horizontal bed is also suitable for these deposits, but less accurate compared with turbidites.

4.2 The TrackDip method

TrackDip is a method initially developed by Basile et al. (2009) to analyse vertical variations of bedding attitude from borehole data or from stratigraphic sections with strike and dip measures. TrackDip divides the dataset from the borehole or stratigraphic section into different intervals with a given thickness, each interval is called a window. Successive windows from large to small scales are created by dividing the entire section first into two windows (largest window scale), then into three windows and this procedure continues until it reaches the smallest window scale (thinnest interval including at least one bedding attitude measurement). For each window, a mean plane is calculated from bedding plane data included in the window. The mean planes may not have a geologic meaning, but allow comparison with neighbouring windows (windows above and below a given window). When two adjacent windows display different mean planes, the tilt between these planes is calculated, by reference to the upper window. These tilts are defined by a tilt axis, a tilt angle, and a dip direction. The TrackDip results are shown in a graph with window size on X-axis and thickness of the entire section on Y-axis. If similar tilts (similar trend axis, angle, and dip direction) are found at different scales, TrackDip is able to detect and follow these tilts from large to small windows, for the same depth intervals (Figs. S1 and S2). Tilts detected from large to small windows are represented in the graph by lines connecting similar tilts found at different scales. Tilts are considered as significant if they can be tracked at least for three successive scales, with significant tilt angles (>4°), and if axis trend variations do not exceed 30° (Basile et al. 2009).

In case of a sudden change in bedding attitude, the tilt may be detected from large to small windows, with the highest tilt angle found at the smallest windows, because the change is local. For example, the highest tilt angle is expected between the smallest windows below and above an angular unconformity. On the contrary, the largest windows may include other tilts considered as noise and thus a lower angle is expected. In case of gradual change in bedding attitude, the tilt between large windows may represent the entire gradual tilt, and thus the tilt angle is expected to correspond to the highest tilt angle. By contrast, the tilt between small windows represents only one increment of tilting, and thus the tilt angle is expected to be lower. More details for the TrackDip method are given in Basile et al. (2009).

4.3 Calculation of the uplift rates of the diapir roof

The amplitude of diapir roof uplif may be estimated using tilt angles within synkinematic strata related to diapir growth. If biostratigraphic data are available within the synkinematic strata, it is possible to give the associated uplift rates of the diapir crest.

Considering a very simple model for flank tilting related to differential displacement between the diapir axis and its sides (Fig. 5), the uplift of the roof may be roughly estimated following equation 1:

$$Us = Ww * \tan \alpha, \qquad (1)$$

where Us is the uplfit of salt after diapir growth (or differential vertical displacement), α the tilt angle in the wedge and Ww the width of the wedge.

Measurement of the tilt angles has to be made where angles are higher (near the salt) because dip angles of upturned strata vary laterally and vertically within the overburden.

This simple equation may be valid only if the diapir is buried, as each increment of salt rising may trigger flanks tilting (Fig. 5). Therefore, uplift rate of the roof and salt flow velocity may be approximately equal, if convective motions within salt are neglected. If salt reaches the surface, salt may flow without tilting the adjacent wedges, thus salt flow velocity may exceed uplift rate of the diapir edge. In this case, this equation may give underestimated results.

5 Structures in the diapir flanks

5.1 Sedimentary wedges and unconformities

The western part of the Bakio bay (Punta de Bakio cliffs) displays a well exposed wedge-shaped structure (Fig. 6a), adjacent to the salt diapir. From east to west, this wedge is composed of the Bakio Breccias Fm., Punta de Bakio and Jata units (Fig. 6b). The Punta de Bakio unit displays several angular unconformities, including its top (Fig. 6b). All units thin eastward and their bedding dips decrease westward. Strata are dipping from 80° toward the SE (overturned strata) in the basal part the Bakio Breccias Fm., to 20° toward the NW in the uppermost part of the Punta de Bakio unit. In the Bakio breccias Fm., four parabreccia beds (s1 to s4) contain slumped clasts with associated folds (Fig. 4). About 700 m away from the diapir, the Jata unit is horizontal on the coastline. To the SW, the Jata unit unconformably overlies the Bakio breccias Fm., showing that the wedge extends toward the SW. Here the Punta de Bakio unit is absent because it pinches out toward the diapir.

Another wedge crops out along the eastern diapir flank. In the lower part of the wedge, the Bakio breccias Fm. unconformably overlies the overturned Bakio marls unit (unconformity u1 in Fig. 4). The angular unconformity indicates an important tilt of the eastern flank, which occurred before deposition of the Bakio breccias Fm. In the upper part of the wedge, the Bakio breccias Fm. and the Sollube unit display a progressive decrease of their bedding dips, from 60° to 20° toward the SE. To the south, the

Sollube unit unconformably overlies the Bakio breccias Fm., showing that this wedge also extends southward.

5.2 The Gaztelugatxe normal fault

A vertical and irregular fault surface, with an apparent vertical offset (Fig. 7a and b), roughly striking NE-SW at present (N048°, 90°) is located on the Gaztelugatxe Island (Fig. 3a), about 1 km NE of Bakio. The footwall, located NW, is composed of the Gaztelugatxe limestones (Fig. 7a and b). The hanging wall, located SE, is composed of the upper part of the Bakio breccias Fm. (Fig. 7a and b). Although, data are lacking north of the present day coastline, mapping suggests that the Bakio breccias Fm. only develops south of this fault. This suggests that this unit pinches out northward and onlap on the Gaztelugatxe fault scarp.

In the hanging wall, the most tilted beds are located adjacent to the major fault scarp and are slightly drag folded showing a normal displacement (Fig. 7b). Two minor normal faults, roughly parallel to the major fault scarp, offset the upper part of the Bakio breccias Fm. (Fig. 7b and d). These faults are synsedimentary faults since their offsets decrease upward and since they are buried by the uppermost part of the Bakio breccias Fm. (Fig. 7d). SE of these faults, three slumped intervals are located at the top of the Bakio breccias Fm. (s5 slumps, Fig. 4) and in the basal part of the Sollube unit (s6 and s7 slumps, Fig. 4). Below these slumps, another slumped bed passes laterally into parabreccias (Fig. 7b and d).

In the footwall, three other secondary normal faults, locally covered by a ferruginous crust, offset the Gaztelugatxe limestones (Fig. 7c). On these minor fault scarps, slickensides give an apparent right-lateral strike-slip displacement with a normal component. Similarly to the major fault, the uppermost part of the Bakio breccias Fm. buried these minor faults (Fig. 7d).

5.3 Minor faults in the Bakio overburden

Several faults (length of fault traces from metre to decametre scale) crop out in the Bakio overburden. The fault network presents three preferential strikes (N030°, N110° and N160°). These faults similarly affect beds, whatever their degree of tilting (from overturned beds to horizontal beds), suggesting that faulting occurred after flank tilting. Moreover, there is neither evidence for lateral facies nor thickness changes, nor associated gravity-driven deposits. This shows that these faults affected beds after their deposition. Similar fault sets affect the whole Albian sedimentary succession, both east (Matxitxako Cape, located about 4 km east of the Bakio diapir) and west of Bakio (from Bakio to Armintza). This means that they are not related to the diapir growth, but to regional deformation postdating the diapir growth.

6 Structural analyses and impact of diapir growth on sedimentation

This section offers a structural interpretation for the major outcrops described in the previous section. Additionally, sedimentological data from two stratigraphic sections studied on either flank of the diapir are discussed.

6.1 Interpretation of the sedimentary wedges and unconformities

Stereographic plots of bedding planes from the western wedge (Fig. 8c and d) give the tilt axis orientation responsible of bedding dip variations in the wedge (Fig. 6). It indicates a progressive tilt with an horizontal axis striking N067°, in the coastline (Fig. 8c) and an axis striking N042° dipping 19°NE, in the second outcrop (Fig. 8d), to the south.

Stereographic plots of bedding planes from the eastern wedge (Fig. 8f and g) show a progressive tilt with an axis striking N009° dipping 03°NE in the coastline (Fig. 8f) and an axis striking N021° dipping 9°N, in the second outcrop (Fig. 8g), to the south.

The map trace of the diapir, with the western and eastern flanks striking NE-SW and N-S, respectively, suggests northward convergence of the flanks (Fig. 8a). Tilting axes calculated from bedding planes in wedges are parallel to the flanks and to the mean diapir axis (Fig. 8b). This demonstrates the creation of wedges and angular unconformities by flanks tilting (u1 to u5, Fig. 4) during diapir growth. Therefore, the Bakio breccias Fm., Sollube and Punta de Bakio units and the base of the Jata unit were deposited during diapir growth.

6.2 Interpretation Gaztelugatxe normal fault

The Gaztelugatxe outcrop has been interpreted as a synsedimentary normal fault scarp (García-Mondéjar and Robador, 1986-1987 ; Robles et al. 1988). The irregular aspect of the fault scarp suggests submarine erosion of the scarp (García-Mondéjar and Robador, 1986-1987; Robles et al. 1988). Fault tion occurred during deposition of the Bakio breccias Fm., since the fault is covered by the uppermost part of this unit (Fig. 7d). This supports the geometry of the Bakio breccias Fm. that seems to pinch out northward, on the fault scarp. This implies uplift in the footwall and deposition of the Bakio breccias Fm. in the hanging wall occurring at the same time. Limestones blocks in breccias and the Gatelugatxe limestones from the footwall exhibit similar facies, thus the Gatelugatxe limestones were the source for breccias, as proposed by García-Mondéjar and Robador (1986-1987). The thick slumped bed changing laterally into parabreccias, in the hanging wall (Fig. 7b) reveals downslope flow transformation toward the SE, consistent with a source from the north, from the Gaztelugatxe limestones.

Reworking related to tectonics, probably from the Gaztelugatxe fault scarp is demonstrated by the size and abundance of olistolithes, up to 1 m thick, in the Bakio breccias Fm. Orthobreccias with thick

olistolithes in the lower part of the Bakio breccias Fm. implies an episode of intense fault activity. On the contrary, thick marly intervals without olistolithes in the upper part of this unit indicate a decrease in fault activity. The onlap of the uppermost part of the Bakio breccias Fm. that buried the fault and the of absence of olitolithes in the overlying Sollube unit confirm the decrease of the fault offset.

The entire structure underwent a late tilting toward the SE (approximately 35° along an axis trending N025°) as beds covering the fault are tilted (Fig. 7d). This tilt may be related to Pyrenean shortening. In order to obtain the orientation of structures during Albian times, the uppermost beds of the structure may be back tilted to horizontal (Fig. 8e). After back tilting of these beds, the three small faults and their slickensides (with an apparent right-lateral strike-slip displacement, prior to back-titling) found top of the footwall, display a normal displacement (Fig. 8e), consistent with the entire structure.

The Bakio diapir and the major Gaztelugatxe fault are aligned together and share comparable trends, respectively N034° and N048° (fig. 8a, b and e). The hanging wall of the Gaztelugatxe fault coincides laterally with the subsiding part of the eastern diapir flank, yet the footwall and the uplifting apex of the diapir may be connected. The Bakio breccias Fm. wedges out toward both the Gaztelugatxe fault and the diapir. This points out that both structures activated at the same time and were connected. Consequently, the diapir probably extends offshore, below the footwall of Gaztelugatxe fault.

6.4 Slumps triggered by the diapir flanks tilting

Numerous slumped intervals can be observed on both sides of the Bakio diapir, in the Punta de Bakio outcrop (western flank) as at Gaztelugatxe (eastern flank).

Measurements of bedding planes attitudes in each slumped interval in the western flank (s1 to s4) have been made to define the slump fold axes. Restored fold axes from the four intervals have similar N-S to NW-SE orientations (Fig. 9a, b, c, d).

In the hanging wall of the Gaztelugatxe fault, direct measurements of slump fold axes have been made. After restoration to horizontal bedding, slump axes trend preferentially E-W in the s5 interval (Fig. 9e) and roughly NE-SW in the s7 slumped interval (Fig. 9f). In the s6 slumped interval, restored slump axes preferentially trend N160° to N180°, with important dispersal of fold hinges (Fig. 9g, h). Such dispersal of fold hinges is linked to the small length and incurved shape of slump folds, with fold terminations almost perpendicular to the central part (Fig. 9i). Measurements of axial planes have been added to provide more information for slumps movements (Fig. 9g), as in Alsop and Marco (2011). Axial planes also show an important dispersal but with a preferential plan striking N027°, 25°NW, indicating eastward verging folds. Some asymmetric slumps indicate an eastward displacement on an eastward dipping slope.

In the Punta de Bakio section, the restored fold axes with N-S to NW-SE orientations for the four intervals (Fig. 9a, b, c, d) parallel the diapir axis (Fig. 4b). Thus, they are interpreted as resulting from destabilization of sediments during flank tilting and sliding toward the west or the NW.

In the hanging wall of the Gaztelugatxe fault, slump axes orientations of the s5 interval suggest a southward-dipping slope, whereas the s6 and s7 interval indicate an eastward and a southeastward-dipping slopes, respectively. This points out changes in orientation of the slope through times. Uplift of the footwall of the Gaztelugatxe fault may produced a southward dipping slope associated with the s5 interval. Since the fault became inactive during deposition of the Monte Sollube unit, other processes may explain the s6 and s7 slumped intervals. Diapir flank tilting may created eastward and southeastward dipping slope producing the s6 and s7 slumped intervals. Uplift to the north may have lasted throughout Late Albian due to diapir growth, even if the fault became inactive. Sediment destabilization from the edge of the Landes Massif, to the north may also explain the s7 slumped interval.

Slump with limestones blocks emanating from the overburden roof, in both flanks attest that the source area for gravity-driven deposits including parabreccias is the diapir roof. This implies the development of a carbonate platform also on top of the Bakio diapir, probably due to shallowing and creation of topographic high on the sea floor.

Paleocurrent directions were measured in sole marks of turbidite beds intercalated with slumps and breccias around the Bakio diapir (in the Sollube, Punta de Bakio and Jata units). Turbidites of the Punta de Bakio unit flowed roughly parallel to the western flank (Fig. 8a). This suggests that the diapir diverted turbidity flows originated from the northern Landes Massif. In the Gaztelugatxe Island, turbidites of the Sollube unit roughly proceeded from the north, with an important dispersion (Fig. 8a). This suggests that turbidity flows emanating from the Landes Massif flowed southward without being influenced by surface relief. The absence of relief at the sea floor confirms that the Gaztelugatxe fault was buried during deposition of the Sollube unit.

Turbidites around the Bakio diapir are interbedded with slumps and parabreccias with limestones blocks, derived from the diapir roof. It implies a multi-source system with turbidity flows emanating from the Landes Massif and breccias with limestones blocks coming from the diapir roof.

6.5 Tilts in the Punta de Bakio wedge using the TrackDip method

The TrackDip method (see 4.2 for details of this method) has been used for the Punta de Bakio stratigraphic section, after a systematic measurement of bedding planes attitudes and position (Fig. S1). It represents 135 bedding measurements in a 277 meters-thick section. The tilts are referenced by a number attributed during TrackDip processing. As all beds are tilted toward the NW, due to wedge tilting, arrows pointing toward the NW indicate steeper beds in the lower windows. By contrast, arrows pointing toward the SE indicate steeper beds in the upper windows. This method enables the detection

of significant geologic structures within the studied section, including unconformities and slumps. Only the tilts with geologic significance are presented in Tables S1 to S4.

The tilt n°1 (Table S1) represents a gradual tilt over the whole studied section, as it shows higher tilts between large windows (see 4.2 for methods details). The tilt axis trends N058°, comparable to the axis previously calculated for the entire wedge (N067°, Fig. 8c). Both northwestward tilt direction and axis orientation are consistent with diapir flank tilting. Therefore, this gradual tilt is interpreted as the main tilt that created the sedimentary wedge.

The tilts n°2, 3, 4 and 5 represent sudden tilts as the maximal tilt angles are found for the smaller window sizes (Table S1). Tilts n° 2 and 4 enable to detect the unconformities u4 and u2, respectively. Tilts n° 3 and 5 may also be related to an unconformity undetected on the field. The tilts n° 1 to 5 have similar rotation axis trend and rotation direction, with the exception of tilt n° 4. Therefore the progressive tilt observed for the whole wedge may be interpreted as the addition of several localized tilts (e.g. tilts n°2 to 5). Tilt n° 4 has an opposite rotation direction, comparatively to the tilts n° 1, 2, 3 and 5 (Fig. S1). Such tilt toward the SE indicate steeper beds in the upper windows. As the tilt n°4 coincides with the base of the s2 slumps, slumps may explain why beds are steeper in the upper windows.

Tilts found at smaller scales display more variations on tilt axis trends. Many changes of orientations are associated with slumped intervals (Fig. S2). In each intervals, tilts axes and measured folded slumps axes are similar. For example, in the s2 slumped interval, tilts (tilts n° 10, 24, 25, 34, 38, 44 and 47, table S2) and slumps (axis trending N016°, Fig. 9b) share similar axes roughly trending NE-SW (Fig. S2a). In the s3 slumped interval, tilts (tilts n° 22, 23, 41, 42, 43 and 46, table S3) trend N-S (Fig. S2a), like the N179° fold axis (Fig. 9c). The change of orientation of tilt axis trends also correlates well with the s2-s3 boundary (Fig. S2a). It may correspond to different fold orientations within the s2 and s3 slumped intervals.

At the smallest window sizes (≤ 10 m), several tilts can be associated as pairs showing opposing dip directions (n° 11 and 18, n°16 and 30, n°19 and 29 and n°20 and 32, table S4), mostly in slumped structures. For example, the n° 11-18 pair displays similar axes (trending N075° and N065°, respectively) with opposing dip directions (Fig. S2b). Opposing dip directions indicate the presence of steeper beds in the interval comprised between these tilts compared to both the upper and lower intervals, because it is related to slumped structures and not directly to diapir flank tilting. These tilts coincide with the s4 slumped interval (Fig. S2b) and have a trend similar to the fold axis of slumps (N037°, Fig. 9d). This is also the case for the n° 16-30 pair (trending N098° and N068°, respectively), which fit on the s1 slumped interval (Fig. S2c). Similar interpretations may explain the pairs of tilts n° 19 and 29 and n° 20 and 32, as both pairs coincide with slumped beds or breccia beds (Fig. S2c).

6.6 Halokinetic sequences

According to Giles and Lawton (2002) and Rowan et al. (2003), angular unconformities in diapir flanks may be interpreted as boundaries of halokinetic sequences. A halokinetic sequence corresponds to a set of strata, locally bounded by unconformities, with gravity-driven deposits in the lower part, and fine-grained deposits in the upper part (Fig. 1c).

In the upper part of the Punta de Bakio stratigraphic section, a 3 m-thick bed of parabreccia is overlain by siliciclastic turbidites (235-244 m, Fig. 10). This package is bounded by the u4 and u5 unconformity and thus may be interpreted as a halokinetic sequence (Fig. 10). Parabreccias may result from doming in the diapir crest, which creates an unstable slope failing periodically (when diapir rising > accumulation rate), while turbidites onlap the diapir (when diapir rising < accumulation rate). Parabreccias are interpreted as relatively proximal slope apron facies since they exhibit large floatstone blocks (up to 1 m wide) with basal scours, and are often non-graded or inversely graded. The sedimentary packages bounded by the u2, u3 and u5 unconformities may also be interpreted as two other halokinetic sequences (Fig. 10).

The TrackDip method suggests that parabreccias and slumps resulted from sliding along the flanks during diapir growth, as these beds display internal structures with tilt axes roughly parallel to the diapir. This suggests that each package of turbidites overlying parabreccias may correspond to halokinetic sequence. Therefore, from 207 to 235 m, 4 additional halokinetic sequences may be defined, even though angular unconformities were not observed (Fig. 10). Below the Punta de Bakio unit, in the upper part of the Bakio breccias Fm., there is no clear repetition of such sedimentary packages (Fig. 4), suggesting the dominance of chaotic gravitational processes. Each parabreccia bed and its overlying marls may correspond to a halokinetic sequence. Stacked parabreccia beds may be interpreted as the result of a high repetition of thin halokinetic sequences, with erosion or reworking of the upper part of the underlying sequences.

Lateral facies variations are found within the halokinetic sequences (Fig. 10). Parabreccias are more abundant and relatively thicker near the diapir (Fig. 10). On the contrary, siliciclastic turbidite facies are thicker away from diapir (Fig. 10). Parabreccias probably wedge out away from the diapir, whereas turbidites onlap on the diapir. These geometries are consistent with slumping and currents directions found in 6.4 section (Fig. 8a), as turbidites proceeded from the North, whereas slumps and parabreccias were triggered by diapir growth.

In the eastern wedge, breccias are almost absent in the Sollube unit (Fig. 4). According to the halokinetic sequences model, the diapir may be buried and gravity-driven deposits are absent, when sedimentation exceeds the diapir rise. As the Sollube unit is up to 1500 m thick, SE of Bakio (Vicente-

Bravo and Robles, 1991a), this suggests high sedimentation rates for this unit. Therefore the Sollube unit probably onlapped on and buried the eastern diapir flank.

6.7 Tertiary shortening

In the Punta de Bakio outcrops, bedding planes in the Jata unit are horizontal, about 700 m away from the diapir. This points out that flanks tilting only affects a limited area around the diapir. Farther SW, more than 2 km away from the diapir, the Jata unit is dipping about 45° westward, indicating that deformation of this unit in this area is due to later Pyrenean folding and not to Bakio diapir uplift.

Mesoscale folds are found around the Bakio diapir with folds axes plunging westward in the western flank and eastward in the eastern flank, respectively, suggesting tilting of these axes by salt rise after Pyrenean folding. Bedding planes in both diapir flanks are overturned and dip toward each other (Fig. 3b). This geometry may be related to slight Tertiary shortening through the diapir.

7. Discussion

This section provides a discussion of the chronology of the Bakio diapir emplacement, its impact on carbonate platform development and the velocity of diapir growth. It finally addresses a discussion on drag folds and drape and halokinetic sequence models.

7.1 Timing and heterogeneities of the Bakio diapir growth

The oldest unit tilted by diapir growth is the Bakio marls unit (Early Albian). The oldest evidence of diapir growth is the u1 unconformity top of the Bakio marls unit (Fig. 4), since there is neither thinning of this unit toward the diapir nor angular unconformities. The u1 unconformity is located where the Gaztelugatxe fault connects to the diapir, thus it is difficult to decipher whether the dip of the Bakio marls unit is due to diapir growth, faulting, or both. The youngest evidence for diapir rising in the western flank, is the unconformity at the base of the Jata unit (Late Albian) overlying the Bakio breccias Fm. (Fig. 4). Therefore the diapir rose at least since deposition of the Bakio breccias Fm. until deposition of base of the Jata unit, i.e. from early Middle Albian (*dentatus* Zone) to Late Albian times.

All tilt axes of the wedges are oblique to the diapir axis and are roughly convergent northward (Fig. 8a). Both diapir flanks are also convergent northward, indicating that the diapir may be a conic, salt-cored antiform. Convergent tilt axes suggest that the diapir acquired this geometry during sedimentation, probably due to diachronic lateral migration of diapir growth.

In the western flank, the source area for gravity-driven deposits including parabreccias is the diapir roof, since slump with limestones blocks derived from the diapir. This implies the development of

a carbonate platform top of the Bakio diapir, probably due to shallowing and creation of topographic high on the sea floor (Fig. 11a). Blocks in breccias exhibit similar facies in both the western diapir flank and the footwall of the Gaztelugatxe fault and the apex of the diapir connected with the footwall of the fault (Fig. 11a). This suggests that a single carbonate platform developed top of the diapir and the footwall of the Gaztelugatxe fault.

In the southern part of the Bakio diapir, only marls and rare thin-bedded grainstones are found as lateral equivalents to breccias, only located in the northern part of the diapir. This suggests higher rates of salt flow to the north, or that salt uplift began earlier in the north. This may explain the conical geometry of the diapir.

In the western flank, angular unconformities within the Punta de Bakio unit and at its top attest that the diapir growth lasted during the Late Albian in this area. In the Gaztelugatxe Island, flows recorded by turbidites of the Sollube unit flowed roughly southward (Fig. 8a), without being driven by any significant relief, indicating that the Gaztelugatxe fault was buried at this time. In the eastern flank, the Sollube unit gravity-driven breccia deposits are almost absent. This suggests that the Sollube unit probably also onlapped on and buried the eastern flank of the diapir (Fig. 11b), as high sedimentation rate exceeded those of diapir rise. Westward migration of salt may be related to differential loading on the initial salt layer, in the eastern flank, produced by the Sollube unit, which is up to 1500 m thick, SE of Bakio (Vicente-Bravo and Robles, 1991a).

Such results, together with the conical geometry of the diapir, suggest diachronic lateral migration of diapir growth. As the diapir and the Gaztelugatxe fault probably initiated together during the early Middle Albian, the diapir is interpreted as a reactive diapir growing in response to the offset of the fault (Fig. 11a). Movements of the Gaztelugatxe fault ceased, while the diapir continued to rise at least until deposition of the Jata unit. Therefore, the Bakio diapir is assumed to have evolved passively, with southwestward migration of salt, during the Late Albian (Fig. 11b)

In the northern part of the Basque Country (including the Bakio area), Agirrezabala (1996) and García-Mondéjar et al. (1996) reported individualization of small troughs filled by Middle Albian coarse-grained gravity-driven deposits (like the Bakio breccias Fm.) separating paleo-highs with Urgonian carbonate platforms. These depocentres were bounded by normal NE-SW and N-S striking faults, suggesting a local E-W or NW-SE extension. The Bakio diapir and the Gaztelugatxe fault are both striking NE-SW and probably initiated in response to the local, E-W to NW-SE extension. In the Basque-Cantabrian Basin, NE-SW faults are commonly interpreted as inherited Hercynian faults (García-Mondéjar, 1996). The Bakio diapir and the Gaztelugatxe fault may be located above a basement fault striking NE-SW and may connect at depth with this latter. Reactivation of this basement fault by local NW-SE extension could have initiated the Bakio diapir growth during early Middle Albian times. Bodego and Agirrezabala (in press) document contemporaneous halokinetic deformations controlled by basement normal faulting in the northeastern margin of the basin. Localization of diapirs above

basement faults has been also reported in the Aquitaine Basin (Canérot et al., 2005). López-Horgue et al. (2010) showed that the Pondra diapir (western Basque-Cantabrian Basin, see Fig. 2) also rose as a reactive diapir, in response to local NW-SE extension, during the Early Albian.

Finally, the Bakio area underwent a limited Tertiary deformation as most of Albian structures were not inverted or only slightly tilted. The latest southeastward tilt of the Gaztelugatxe fault may be linked to Pyrenean shortening. Despite the Tertiary shortening, Albian structures remained first order structures.

7.2 Uplift rates of the diapir roof

In order to determine uplift rates of the diapir roof, an estimation of the entire Albian tilt angle is needed for the wedges (as explain in the Methods section). For the Bakio diapir, the entire tilt angle of each wedge may result of a combination of Albian tilts and Pyrenean shortening. The addition of angles between unconformities and the underlying tilted sediments have been used, rather than the entire tilt angle for each wedge, in order to exclude possible Pyrenean tilt. Assuming that sedimentary surfaces were initially close to horizontal, addition of angles is supposed to correspond to tilts that occurred during sedimentation. Since schistosity is absent in the study area, these angles were probably not significantly affected by Pyrenean shortening.

However, sedimentation surfaces were not horizontal in all cases, because several surfaces around the diapir correspond to scars resulting from sliding along dipping slopes. It is not possible to accurately determine the initial dip of these erosional surfaces. Because of their transport as flows, turbidites are supposed to be deposited on sub-horizontal slopes. On the contrary, mass deposits can develop on relatively dipping slopes (not higher than 10° in submarine environments). Therefore, the estimation of the tilt angle between unconformity and underlying tilted beds is better when unconformities are overlaid by turbidites rather than mass deposits. When sediments deposited over angular unconformities are mass deposits, the angle between erosional surface and underlying tilted sediments may be underestimated (up to 10°).

In the eastern flank, the s1 unconformity represents a 54° tilt between the underlying and overlying sediments. As the overlying sediments are composed of breccias, the tilt angle may be underestimated. In the western flank, there are four angular unconformities developed in the Punta de Bakio wedge (u2, u3, u4 and u5, from 234 to 267 m). The measured tilt for u2 and u3 unconformities are 7° and 17°, respectively. It was not possible to acquire measurement of bedding planes orientation between the u4 and u5 unconformities, but comparison from below u4 and above u5 unconformities gives an estimated angle of 14°. The addition of these angles gives a total angle of 38°. This value is similar to the tilt n° 1, with a tilt angle of 35° (Fig. S1).

As explain in the Methods section, calculation of tilt angles allows estimation of the diapir roof uplift rate, since biostratigraphic data are available within synkinematic strata. In this study, only a range from two extreme values can be estimated due to poor biostratigraphic constraints. In the eastern flank, the tilt (54°) occurred after deposition of the Bakio marls unit (Early Albian) and before deposition of the Bakio breccias Fm. (*dentatus* Zone) in a time span about 1 Myr or less (based on the geologic time-scale of Ogg and Hinnov, 2012). In the western flank, the tilt (38°) occurred during deposition of the upper part of the Bakio breccias Fm. and of the Punta de Bakio unit, and thus may have lasted 1.5 Myr. It gives a range of the diapir roof uplift rates from 0.36 to 6.8 mm a^{-1}. As explained above in the Methods section, such values may have different significations. If the diapir is buried, the values correspond to the diapir roof uplift rate and approximately equals the salt flow velocity. If salt reaches the surface, these values represent only the uplift rate of the edge of the diapir as the salt flow velocity may be higher. In the Bakio area, there is no evidence of emergence of salt during sedimentation, as salt tongues or Triassic clasts within deposits from the overburden are absent. Therefore, the range of uplift rates and salt flow velocity may be approximately equal.

In the literature, rates of salt flow and uplift are estimated using various methods. For example, Talbot and Jarvis (1984) and Talbot and Aftabi (2004) postulated simple hypotheses in which, the salt flow velocity is only driven by the weight and the density of the overburden. These authors provided an equation for calculation of salt flow velocity. They found values from 82 mm a^{-1} (Talbot and Aftabi, 2004) to 170 mm a^{-1} (Talbot and Jarvis, 1984) for salt flow velocity for two diapirs in Iran. Frumkin (1996) used radiochronologic dating on Neogene-Pleistocene uplifting salt from the Dead Sea and obtained values from 6 to 7 mm a^{-1}, for salt flow velocity. Pirazzoli et al. (2004) used radiochronologic dating on uplifted coral-reef Quaternary marine terraces, developed top of diapirs from the Iranian coast and calculated an uplift of diapir roof equal to 2 mm a^{-1}. Aftabi et al. (2010) estimated salt flow velocity from 511 mm a^{-1}, also from another Iranian diapir, using InSAR mapping.

The large discrepancies of the values, different methods, time-scales of observations and geologic contexts make the comparison between these studies difficult. The different time-scales of observations are an important unsolved problem to compare these data. For example, rates of salt flow recorded with InSAR represent punctual measurements (14 years for Aftabi et al., 2010) compared to the results of this study (about 1.5 Myr). Since salt flow velocity is considered to change rapidly over short periods of time, as proposed by the halokinetic sequence model (Giles and Lawton, 2002; Rowan et al. 2003), short-term results may represent pulses of salt growth whereas long term results are likely to be time-average.

If this problem of time-scales of observations is neglected, a trend of diapirs with fast (> 10 mm a^{-1}) and slow (< 10 mm a^{-1}) rates of salt flow may be defined. The results of this paper are similar to those of Frumkin (1996) and Pirazzoli et al. (2004) with relatively slow motions compared to other studies

(Talbot and Jarvis, 1984; Talbot and Aftabi, 2004; Aftabi et al., 2010). It may be argued that the geologic context drastically influence rates of salt flow as Iranian diapirs studied by Talbot and Jarvis (1984), Talbot and Aftabi (2004) and by Aftabi et al. (2010) show faster rates of salt flow. It may be related to important shortening occurring in the Zagros Montain. On the contrary, diapirs with slow rates of salt flow (or diapir roof uplift) from Bakio (this study), the Dead Sea (Frumkin, 1996) and the Persian Gulf (Pirazzoli et al., 2004) developed on subsiding areas. Moreover, in diapirs with fast rate of salt flow (Talbot and Jarvis, 1984; Talbot and Aftabi, 2004; Aftabi et al., 2010), salt reaches surface and flows upward rapidly, which was probably not the case for the Bakio diapir. In addition, diapirs with fast rates of salt flow are developed in continental area and are submitted to dissolution by rainfall as salt emerged at surface. Dissolution of salt top of diapirs may considerably increase salt velocity flow. The comparison between these studies suggests that diapirs with fast (> 10 mm a^{-1}) salt velocity flow may develop in continental areas with important shortening, with salt emerging at surface, submitted to dissolution by rainfall. On the contrary, diapirs with slow (< 10 mm a^{-1}) salt velocity flow rather develop in subsiding areas and are probably buried. This was probably the case of the Bakio diapir.

In this study, sparse biostratigraphic data do not allow an accurate estimation of velocity of diapir growth. However, the estimation of uplift rate of the diapir roof, constrained by accurate biostratigraphic data may be a novel approach to constraining rates of halokinesis in other diapirs. This method may be useful to demonstrate changes of rate over time and thus demonstrate relative growth rates between halokinetic sequences. Further investigation is still needed around the Bakio diapir to provide better biostratigraphic data, in order to refine the range of velocity of diapir growth.

7.3 Drag folds *versus* drape folds

Structures in the overburden of the Bakio diapir are mainly angular unconformities, sedimentary wedges and slumps, indicating that deformation occurred at sea floor. Tilt axes of the slumps and wedges are parallel to the diapir trend (Fig. 8) and thus are clearly related to diapir rise. Major angular unconformities indicate that wedge tilting affected the sea floor (Fig. 6). The TrackDip results demonstrate that flank tilting with deformation occurring at the sea floor may explain most of the tilts within the western wedge. Slumps and parabreccias imply a dipping slope and significant relief related to the diapir apex. Such relief on the sea floor is confirmed by paleocurrent directions around the Bakio diapir, as the turbidity currents were driven by the diapir relief (Fig. 8a).

By contrast, most of faults affecting the overburden are related to regional tectonics and not to diapir growth. This suggests the absence of significant brittle deformation linked to diapir growth in the overburden. This suggests that most of deformation caused by diapir growth occurred in poorly consolidated sediments.

These results are not compatible with drag fold model (Jackson and Talbot, 1991; Jackson et al., 1994; Alsop, 1996; Alsop et al., 1995 and 2000), as this model cannot explain different upturn degrees

separated by unconformities in the wedges. Wedges with overturned beds, angular unconformities and slumps around the Bakio diapir are similar to those described by Giles and Lawton (2002) and Rowan et al. (2003) from the El Papalote diapir (Fig. 1c). Diapir growth related deformation occurred near or at surface and not at depth, therefore this support the drape fold model.

7.4 Halokinetic sequences

In the western diapir flank, the data show that slumps are triggered by tilting the sea floor during diapir growth. These slumps are composed of folded grainstones and orthobreccia beds and limestones blocks within a marly matrix. Parabreccia beds display the same types of clasts and matrix, which suggests a genetic relationship with the slumps. Downslope flow transformation of slump to parabreccias may be inferred, regarding the similitude in clasts and matrix. Therefore, parabreccias are assumed to result of mud-supported debris flows derived from the diapir roof.

Slumps and parabreccias are interbedded with turbidites and marls indicating episodic production of slumps and mud-supported debris flows by slope failure. Paleocurrent and slumping directions around the Bakio diapir indicate multi-source system with slumps and parabreccias triggered by diapir rise, and turbidity flows coming from the Landes Massif (Fig. 8a). Parabreccias alternating with turbidites suggest episodic changes in direction of sedimentary transport, together with a change of sedimentary processes. Packages composed of parabreccias in the lower part and thin-bedded fine-grained turbidites in the upper part, separated by unconformites, can be defined in the Punta de Bakio stratigraphic section (Fig. 10). These packages have similar thickness from 4 to 12 m (Fig. 10), indicating cyclic repetition of slope failure.

Such sedimentary packages, composed of parabreccias alternating with turbidites, are interpreted as halokinetic sequences since diapir growth causes cyclic repetition of slope failure as described by Giles and Lawton (2002) and Rowan et al. (2003). According to the halokinetic sequences model, when rates of sedimentation are low relative to those of salt rise, tilted sediments on diapir flanks reach slope failure, slide, and gravity-driven deposits develop at the toe of the slope. When rates of sedimentation are high relative to those of salt rise, onlap on the flanks prevent sediments to reach slope failure. In the Punta de Bakio stratigraphic section, doming in the diapir crest (when diapir rising velocity > accumulation rate) triggered mud-supported debris flows, forming the lower part of halokinetic sequences. Turbidites constituting the upper part of halokinetic sequences, came from the north and onlapped the slope previously formed (when diapir rising < accumulation rate). Such interaction between gravity-driven deposits triggered by diapir growth and regional deposits may be found in other diapirs, if the input of regional sediments is independent of diapir growth.

The halokinetic sequences from the western wedge are characterized by low-angle unconformities (u2-u6) and are associated with broad zone of drape folding (about 700 m), thus may correspond to wedge halokinetic sequence as described by Giles and Rowan (2012). In the eastern flank, the u1 major unconformity shows a high tilt angle (54°) and may correspond to the basal part of a hook halokinetic sequence. According to Giles and Rowan (2012), wedge and hook halokinetic sequences are found when sedimentation rates are superior and inferior to the rates of diapir rise, respectively. In the Basque Trough, the onset of siliciclastic turbidite deposition, during the Late Albian, coincide with an increase of sedimentation rates since the Black Flysch units may be 7000 m thick (Martín-Chivelet et al., 2002; García-Mondéjar et al., 2004). Therefore, prior to the Late Albian, the Bakio diapir growth probably produced hook halokinetic sequences, as suggested by the u1 unconformity and the lower sedimentation rates. Then, during the Late Albian, diapir growth produced wedge halokinetic sequences because of the high sedimentation rates associated with siliciclastic turbidites deposition. It demonstrates that the hook and wedge halokinetic sequences model is relevant for the Bakio diapir.

8 Conclusions

The Bakio diapir and its overburden provide new important data from an outcrop case with syn-diapir growth strata, well-exposed sedimentary wedges, major unconformities and upturned strata. The study of this diapir brings new key elements for: (i) local understanding of the timing of diapir growth, which was previously unknown, (ii) insights for the drape folds and (iii) halokinetic sequences models and offers (iv) a new method for estimating rates of diapir growth. These new key elements are listed below:

(i) The Bakio diapir probably initiated during the early Middle Albian as a reactive diapir in response to the offset of the Gaztelugatxe fault, and evolved passively from Middle to Late Albian times. Rate of diapir growth was variable through times, as salt probably started rising north of Bakio during the early Middle Albian and then preferentially southward, when movements of the Gaztelugatxe fault ceased during the Late Albian. The range of rates of salt rise is estimated from 0.36 to 6.8 mm a^{-1} during the passive stage of the diapir. During the Tertiary shortening, linked to the Pyrenean orogeny, the diapir was probably gently reactivated.

(ii) Structural data and analysis with the TrackDip method show deformation at the sea floor and absence of shear generated by salt flow at depth in the overburden. This demonstrates the relevance of the drape fold model for the Bakio diapir, yet the drag fold model is not suitable. This conclusion may be valid for for other diapir beyond the La Popa Basin. Further studies on other diapirs are still needed to confirm this hypothesis.

(iii) Unconformity-bounded sedimentary packages composed of breccias in the lower part and turbidites interbedded with marls in the upper part are interpreted as wedge halokinetic sequences. This points out that the halokinetic sequences model described previously only for the La Popa basin, is relevant for

the Bakio diapir and thus probably for other passive diapirs. In addition, this study illustrates how turbidites produced by flows emanating from another source are interbedded with gravity-driven deposits triggered by diapir growth. Such interplay between regional sediments and gravity-driven deposits may be found in other diapir, where the source of regional sediments is independent of diapir growth.

(iv) This work offers a new simple method for estimating the velocity of diapir growth, based on the estimation of the velocity of diapir flank tilting by dating synkinematic strata with biostratigraphic data. This method may be useful for other outcrop studies where biostratigraphic data are available.

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References:

Aftabi, P., Roustaie, M., Alsop, G. I., and Talbot, C. J. (2010). InSAR mapping and modelling of an active Iranian salt extrusion. Journal of the Geological Society, 167, 155-170.

Agirrezabala, L.M., and García-Mondéjar, J. (1989). Evolución tectosedimentaria de la plataforma urgoniana entre Cabo Ogoño e Itziar (Aptiense-Albiense superior, Región vasco-cantabrica oriental). In: Syrnposium, XIZ Congreso Español de Sedimentologia (Ed. by S. Robles, J. García-Mondéjar and A. Garrote), 11-20, UPV & EVE, Bilbao.

Agirrezabala, L.M. 1996. El Aptiense-Albiense del Anticlinorio Nor-Vizcaino entre Gernika y Azpeitia. PhD Tesis, Euskal Herriko Unibertsitatea, Bilbo, 429 pp.

Ábalos, B., Alkorta, A., and Iribar, V. (2008). Geological and isotopic constraints on the structure of the Bilbao Anticlinorium (Basque-Cantabrian Basin, north Spain). Journal of Structural Geology, 30, 1354-1367.

Alsop, G. I., Jenkins, G. and Davison, I. (1995). A preliminary study of drag zone geometry adjacent to salt diapirs, in Salt, sediment and hydrocarbons. Gulf Coast Section, SEPM 16th Annual Research Conference, p. 1–9.

Alsop, G. I. (1996). Physical modelling of fold and fracture geometries associated with salt diapirism, in Alsop G. I., Blundell D. J., and Davison I., eds., Salt tectonics. Geological Society (London) Special Publication 100, p. 227–241.

Alsop, G. I., Brown, J. P., Davison, I. and Gibling, M. R. (2000). The geometry of drag zones adjacent to salt diapirs. Journal of the Geological Society (London) v. 157, p. 1019–1029.

Alsop, G. I. and Marco, S. (2011). Soft-sediment deformation within seismogenic slumps of the Dead Sea Basin. Journal of Structural Geology, 33, 433-457.

Banham, S.G. and Mountney, N.P., 2013. Evolution of fluvial systems in salt-walled minibasins: a review and new insights. Sedimentary Geology.

Banham, S.G. and Mountney, N.P., 2013. Climatic versus halokinetic control on sedimentation in a dryland fluvial succession: Triassic Moenkopi Formation, Utah, USA. Sedimentology.

Banham, S.G. and Mountney, N.P., 2013. Controls on fluvial sedimentary architecture and sediment-fill state in salt-walled mini-basins: Triassic Moenkopi Formation, Salt Anticline Region, SE Utah, USA. Basin Research.

Barton, D. C. (1933). Mechanics of formation of salt domes with special reference to Gulf Coast salt domes of Texas and Louisiana. American Association of Petroleum Geologists Bulletin, v. 17, 1025–1083.

Basile, C., Pecher, A., Corazzi, M., Odonne, F., Maillard, A., Debroas, E.J. and Callot, P. (2009). TrackDip: A multi-scale processing of dipmeter data – method, tests, and field example for 3-D description of gravity-driven deformations in the Eocene foreland basin of Ainsa, Spain. Marine and Petroleum Geology, 26, 738–751.

Bodego, A. and Agirrezabala, L.M., (in press). Syn-depositional thin- and thick-skinned extensional tectonics in the mid Cretaceous Lasarte sub-basin, western Pyrenees. Basin Research (2013) 25, doi: 10.1111/bre.12017

Canérot, J., Hudec, M., and Rockenbauch, K. (2005). Mesozoic diapirism in the Pyrenean orogen: Salt tectonics on a transform plate boundary. American Association of Petroleum Geologists Bulletin, 89, pp. 211–229.

Davison, I., Alsop, I., Birch, P., Elders, C., Evans, N., Nicholson, H., Rorison, P., Wade, D., Woodward, J., Young, M., (2000). Geometry and late-stage structural evolution of Central Graben salt diapirs, North Sea. Marine and Petroleum Geology 17, 499–522.

Frumkin, A. (1996). Uplift rate relative to base-levels of a salt diapir (Dead Sea basin, Israel) as indicated by cave levels. Geological Society, London, Special Publications, 100, 41-47.

García-Mondéjar, J. and Robador A. (1986-1987). Sedimentación y paleogeografía del Complejo Urgoniano (Aptiense-Albiense) en el area de Bermeo (region Vasco-Cantábrica septentional). Acta Geológica Hispánica, t. 21-22, pp. 411-418.

García-Mondéjar, J. (1996). Plate reconstruction of the Bay of Biscay. Geology, v. 24, no. 7, 635-638.

García-Mondéjar, J., Agirrezabala, L.M., Aranburu, A., Fernández-Mendiola, P.A., Gómez-Pérez, I., López-Horgue, M.A., Rosales, I., (1996). The Aptian-Albian tectonic pattern of the Basque-Cantabrian Basin (northern Spain). Geological Journal 31, 13-45.

García-Mondéjar, J., Fernández-Mendiola, P.A., Agirrezabala, L.M., Aranburu, A., López-Horgue, M.A., Iriarte, E., Martinez de Rituerto, S. (2004). El Aptiense-Albiense de la Cuenca Vasco-Cantábrica. En: Geología de España, SGE-IGME, pp. 291-296).

Giles, K. A., and Lawton, T. F. (2002). Halokinetic sequence stratigraphy adjacent to the El Papalote diapir, northeastern Mexico. American Association of Petroleum Geologists Bulletin, v. 86, 823–840.

Gómez, M., Vergés, J., and Riaza, C. (2002). Inversion tectonics of the northern margin of the Basque Cantabrian Basin. Bulletin de la Société Géologique de France, v. 173, no 5, 449-459.

Handy, M.R., Schmid, S.M., Bousquet, R., Kissling, E., Bernoulli, D (2010). Reconciling plate-tectonic reconstructions of Alpine Tethys with the geological-geophysical record of spreading and subduction in the Alps. Earth Science Review. 102, 121-158.

Hudec, M. R., Jackson, M. P., and Schultz-Ela, D. D. (2009). The paradox of minibasin subsidence into salt: Clues to the evolution of crustal basins. Geological Society of America Bulletin, 121, 201-221.

Ings, S. J., and Beaumont, C. (2010). Shortening viscous pressure ridges, a solution to the enigma of initiating salt 'withdrawal'minibasins. Geology, 38, 339-342.

Jackson, M. P. A., and Talbot, C. J., (1991). A glossary of salt tectonics. University of Texas at Austin, Bureau of Economic Geology Geological Circular No. 91-4, 44.

Jackson, M. P. A., Vendeville, B. C., and Schultz-Ela, D. D. (1994). Structural dynamics of salt systems. Annual Review of Earth and Planetary Sciences, v. 22, 93–117.

Jackson, M. P. A., Shultz-Ela, D.D., Hudec, M.R., Watson, I.A. and Porter,M.L. (1998). Structure and evolution of Upheaval Dome: A pinched-off salt diapir. Geological Society of America Bulletin, vol. 110, No. 12, pp. 1547-1573.

López-Horgue, M.A., Owen, H.G., Aranburu, A., Fernández-Mendiola, P.A., García-Mondéjar, J. (2009). Early late Albian (Cretaceous) of the central region of the Basque-Cantabrian Basin, northern Spain: biostratigraphy based on ammonites and orbitolinids. Cretaceous Research 30, 385-400.

López-Horgue, M.A., Iriarte, E., Schröder, S., Fernández-Mendiola, P.A., Caline, B., Corneyllie, H., Frémont, J., Sudrie, M. and Zerti, S. (2010). Structurally controlled hydrothermal dolomites in Albian carbonates of the Ason valley, Basque Cantabrian Basin, northern Spain. Marine and Petroleum Geology, 27, 1069–1092.

Martín-Chivelet, J., et al. (2002). Cretaceous, in The Geology of Spain, edited by W. Gibbons and T. Moreno, pp. 255 – 292, Geological Society of London.

Mutti, E. and Davoli G. (1992). Turbidite Sandstones. Agip, Istituto di geologia, Università di Parma, 275 p.

Nelson, T.H. (1991). Salt tectonics and listric normal faulting. In: A. Salvador (Editor), The Gulf of Mexico Basin (The Geology of North America, J.). Geological Society of America, Boulder, CO, 73-89.

Ogg, G.J. and Hinnov, L.A. (2012). Cretaceous. In: The geologic time scale, Elsevier, pp. 793-853.

Olivet, J.-L. (1996). Kinematics of the Iberian plate. Bulletin des Centres de Recherches Exploration-Production Elf Aquitaine, v. 20, 131-195.

Pirazzoli, P. A., Reyss, J. L., Fontugne, M., Haghipour, A., Hilgers, A., Kasper, H. U., Nazari, H., Preusser, F. and Radtke, U. (2004). Quaternary coral-reef terraces from Kish and Qeshm Islands, Persian Gulf: new radiometric ages and tectonic implications. Quaternary International, 120, 15-27.

Rat, P. (1988). The Basque-Cantabrian Basin between the Iberian and the European Plates. Some facts but still many problems. Revista de la Sociedad de Geologia de España, 1, (3-4).

Robles, S., Pujalte, V., and García-Mondéjar, J. (1988). Evolución de los sistemas sedimentarios del margen continental cantábrico durante el Albiense y Cenomaniense, en la transversal del litoral vizcaino. Revista de la Sociedad de Geologia de España, 1(3-4).

Rosenbaum, G., Lister, G. S., and Duboz, C. (2002). Relative motions of Africa, Iberia and Europe during Alpine orogeny. Tectonophysics, 359, 117-129.

Rowan, M.G., Lawton, T.F., Giles, K.A. and Ratliff, R.A. (2003). Near-salt deformation in La Popa basin, Mexico, and the northern Gulf of Mexico: a general model for passive diapirism. American Association of Petroleum Geologists Bulletin, 87:733–756.

Schultz-Ela, D. D. (2003). Origin of ‘‘drag’’ folds bordering salt diapirs. American Association of Petroleum Geologists Bulletin, v. 87, p. 757–780.

Sibuet, J.C., Srivastava, S.P., and Spakman, W. (2004). Pyrenean orogeny and plate kinematics. Journal of Geophysical Research, Washington, vol. 109.

Stewart, S.A. (2006). Implications of passive salt diapir kinematics for reservoir segmentation and radial and concentric faults. Marine and Petroleum Geology. 23, 843–853.

Talbot, C. J., & Jarvis, R. J. (1984). Age, budget and dynamics of an active salt extrusion in Iran. Journal of Structural Geology, 6, 521-533.

Talbot, C., & Aftabi, P. (2004). Geology and models of salt extrusion at Qum Kuh, central Iran. Journal of the Geological Society, 161, 321-334.

Vendeville, B. C., and Jackson, M. P. A., (1992). The rise of diapirs during thin-skinned extension: Marine and Petroleum Geology, v. 9, p. 331–353.

Vendeville, B. C., and Jackson, M. P. A. (1993). Some dogmas in salt tectonics challenged by modeling (abs.). American Association of Petroleum Geologists Hedberg International Research Conference on Salt Tectonics, Bath, England, September 1993, p. 263–265.

Vicente-Bravo, J. C., and Robles, S. (1991a). Geometría y modelo deposicional de la secuencia Sollube del Flysch Negro (Albiense medio, norte de Bizkaia). Geogaceta, 10, 69-72.

Vicente-Bravo, J. C., and Robles, S. (1991b). Caracterización de las facies de la transición canal-lobulo en la secencia Jata del Flysch Negro (Albiense Superior, Norte de Vizcaya). Geogaceta, 10, 72-75.

Vicente-Bravo, J. C. and Robles, S., (1995). Large-scale mesotopographic bedforms from the Albian Black Flysch, northern Spain: characterization, setting and comparisons with recent analogues, in Pickering, K.T., Hiscott, R.N., Kenyon, N.H., Ricci Lucchi, F., and Smith, R.D.A., eds., Atlas of Deep Water Environments: Architectural Style in Turbidite Systems: London, Chapman & Hall, p. 216–226.

Voort, H.B. (1963). Zum flysch problem in den West-pyrenäen. Geologische Rundschau. 53, 220-233.

Weijermars, R., Jackson, M. P. A., and Vendeville, B. C. (1993). Rheological and tectonic modelling of salt provinces. Tectonophysics, v. 217, p. 143–174.

Wiedmann, J. and Boess, J. (1984). Ammoniten funde aus der Biskaya-synkline (Nordspanien)-Kreidegliederung und Alter des Kreide-Vulkanismus. Eclogae Geologicae Helvetiae, 77: 483-510.

Figures caption

Figure 1. Models for diapir growth and overburden deformation. a: active, passive and reactive diapir growth models (modified from Jackson et al., 1994). b: drag fold model (modified from Alsop et al., 2000). c: drape folds model (modified from Giles and Lawton, 2002).

Figure 2. Geologic map of the Basque-Cantabrian Basin (simplified from Ábalos et al., 2008). BCB: Basque-Cantabrian Basin. BT: Basque Trough, Uf: Ubierna fault, Vf : Villaro fault, Bf: Bilbao fault, Gf:

Gernika fault, Lf: Leitza fault, Pf: Pamplona fault, Ld: Larredo diapir, Pd: Pondra diapir, Md: Mena diapir, Gd: Gernika diapir, E-Ld: Estella-Lizarra diapir, Sd: Salinas de Oro diapir, Gud: Gulina diapir. The dotted line indicates the possible location of the edge of the Landes Massif, currently offshore.

Figure 3. a. Geologic map of the Bakio area and, b. Cross-section through the Bakio diapir, based on new field work and data from Robles et al. (1988b and 1989), García-Mondéjar and Robador (1986-1987), Pujalte et al. (1986-1987), Vicente-Bravo and Robles (1991a and b). The dotted line indicates the location of the cross-section.

Figure 4. Stratigraphic sections on the Bakio diapir flanks and their possible correlations based on biostratigraphic data from Wiedmann and Boess (1984), García-Mondéjar and Robador (1986-1987), Robles et al. (1988) and López-Horgue et al. (2009). The Gaztelugatxe limestones are not represented here. These limestones crop out NE of Bakio, in the footwall of the Gaztelugatxe fault and are probably a lateral equivalent of the Bakio breccias Fm. As the Bakio breccias Fm. to the west has the same facies than the upper part of the Bakio breccias Fm. to the east, both part are probably lateral equivalent. The Punta de Bakio and the Jata units are probably the lateral equivalent of the Sollube unit. Unconformities are labeled from u1 to u6 and slumped intervals from s1 to s7, depending on their stratigraphic location.

Figure 5. Simple model of salt rise used to estimate salt height from the dip angle along the flanks, where Us is the height of salt after diapir growth (or differential vertical displacement), α the tilt angle in the wedge and Ww the width of the wedge.

Figure 6. a. Westward view of the Punta de Bakio sedimentary wedge. b. Interpretation of the wedge, composed from bottom to top of the Bakio breccias Fm., Punta de Bakio and Jata units. These units exhibit a westward decreasing dip and the Punta de Bakio unit displays four angular unconformities (u3 to u6, Fig. 5). Near the diapir contact, beds are overturned (about 10 metres to the east of the picture).

Figure 7. a. Aerial view of Gaztelugatxe fault from the SW, with the Gaztelugatxe unit to the NW (footwall) and the upper part of the Bakio breccias Fm. to the SE (hanging wall). b. Detail of the aerial view of the Gaztelugatxe fault from the SW illustrating the irregular fault scarp developed in the Gaztelugatxe unit (GL) and two secondary faults developed in the upper part of the Bakio breccias Fm. (Bb Fm.). S0: bedding planes, m: marls, b: breccias, s: slumps. c. View of minor faults top of the footwall, offsetting the Gaztelugatxe unit (GL). Marls (m) and breccias (b) of the upper part of the Bakio breccias Fm. (Bb Fm.) cover the southernmost fault. Slickensides developed on these faults planes. After tilting to syn-sedimentary orientation by rotating bedding planes covering the faults, these slickensides give a normal offset (Fig. 4e). d. Interpretation of the entire outcrop.

Figure 8. a. Geologic map of the Bakio area. Black rose diagrams correspond to turbidites currents directions (20 measurements in the Monte Sollube unit, east of Bakio; 8 and 6 measurements in the Punta de Bakio and Jata units, respectively, west of Bakio; short classes include 2 measurements and longer classes include 6 measurements) and white arrows correspond to sliding directions of the s1 to s7 slumps. b, c, d, f, g. Stereographic plots (lower-hemisphere equal-area stereonet) of the mean axis of the

diapir (b.) and of the tilt axes (stars) calculated from bedding planes orientations (points, same colors as in map caption) within the wedges along both flanks (c, d, f, g, h). e: Stereographic plot (lower-hemisphere equal-area stereonet) showing the Gaztelugatxe major fault (red plane) and the minor faults top of the footwall of the Gaztelugatxe fault and their slickensides (black planes and circles) restored to syn-sedimentary orientation by tilting bedding planes covering the faults, to horizontal.

Figure 9. Stereographic plots (lower-hemisphere equal-area stereonet) of bedding planes (s1 to s4) and of slumps axes (s5 to s7) within the s1 to s7 slumped intervals from the Punta de Bakio and the Bakio-Gaztelugatxe stratigraphic section (Fig. 5). a, b, c, d. Bedding planes within the s1 to s4 slumped intervals, black stars correspond to the pole of the great circles calculated from bedding planes orientations and indicate slumps axes. For the s1 fold axis, the blue star corresponds to fold axis, directly measured on the outcrop. e, f, g. Slumps folds axes directly measured on the outcrop for the s5 to s7 slumped intervals. For the s6 slumped interval, the best plan calculated from 41 measurements of axial planes has been added. Each plot of bedding planes is restored to syn-sedimentary orientation by tilting each data set with overlying bedding planes to horizontal. The number of data and S0 used for restoration to horizontal is indicating below each plot. h. Rose diagram of restored folds axis within the s6 interval. The preferential trend, striking N-S is consistent with a displacement toward ENE. i. Bended slump hinge of the s6 slumped interval.

Figure 10. Top of the Punta de Bakio stratigraphic section illustrating lateral variations, with thinning toward the diapir and angular unconformities (u to u5, Fig. 5). Below the u3 unconformity, the cyclic patterns of sedimentary packages with parabreccias in the lower part and with turbidites in the upper part suggest that other halokinetic sequences may be present, even if unconformities are not visible.

Figure 11. Schematic 3D diagrams of the Bakio diapir during early Middle Albian (a) and Late Albian (b). a. Initiation of the Bakio diapir as a reactive diapir in response to the offset of the Gaztelugatxe fault during early Middle Albian. b. Passive stage of diapir growth during Late Albian.

Supporting information

Additional Supporting information may be found in the online version of this article :

Figure S1. a: Bedding planes attitudes from the Punta de Bakio stratigraphic section and results from the TrackDip method (tilts for window size less than 10 m are not shown, see Fig. S2). The two first columns represent the dip direction and angle of bedding planes (direction and angle in X-axis), respectively. The last columns shows the results of the TrackDip method (with window size in X-axis, logarithmic scale) over the entire section). The Y-axis for all columns correspond to the stratigraphic thickness. In the last column, each tilt is identified by a grey line giving its position (Y-axis) for each windows size (X-axis). Arrows indicate the downward tilt direction and perpendicular lines give the axis rotation trend; length

of the arrows is proportional with the rotation angle (10° rotation arrow for reference at the top). Vertical grey lines illustrate the investigated windows for each selected tilt.

Figure S2. Details of several tilts that coincides on sedimentary structures as slumps and unconformities. a. Details of the tilt n°4 and of other tilts at smaller window sizes fitting with the s2 and s3 slumped intervals. b. Details of tilts coinciding with the s4 slumped interval. c. Details of pair of tilts that fit on sedimentary structures as the s1 and s4 slumped intervals and on the u4 unconformity.

Table S1. Results of the TrackDip method showing the major tilts for the largest window sizes. The tilt n°1 represents a gradual tilt over the whole studied section and is interpreted as the main tilt that created the sedimentary wedge. The tilts n°2, 3, and 5 represent sudden tilts that coincide with angular unconformities. The tilt n°2 coincides with the s2 slumps.

Table S2. Results of the TrackDip method showing the tilts related to the s2 slumped interval.

Table S3. Results of the TrackDip method showing the tilts related to the s3 slumped interval.

Table S4. Results of the TrackDip method showing pair of tilts with opposing dip directions, related to small slumped structures.