Mid-depth calcareous contourites in the latest Cretaceous of Caravaca (Subbetic Zone, SE Spain). Origin and palaeohydrological significance

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Abstract

Deep marine carbonates of Late Campanian to Early Maastrichtian age that crop out in the Subbetic Zone near Caravaca (SE Spain) contain a thick succession of dm-scale levels of calcareous contourites, alternating with fine-grained pelagites/hemipelagites. These contourites, characterised by an abundance and variety of traction structures, internal erosive surfaces and inverse and normal grading at various scales, were interpreted as having been deposited under the influence of relatively deep ocean currents. Based on these contourites, a new facies model is proposed.

The subsurface currents that generated the contourites of Caravaca were probably related to the broad circumglobal, equatorial current system, the strongest oceanic feature of Cretaceous times. These deposits were formed in the mid-depth (200–600 m), hemipelagic environments at the ancient southern margin of Iberia. This palaeogeographic setting was susceptible to the effects of these currents because of its position close to the narrowest oceanic passage, through which the broad equatorial current system flowed in the westernmost area of the Tethys Seaway. Regional uplift, related to the onset of convergence between Iberia and Africa, probably favoured the generation of the contourites during the Late Campanian to the Early Maastrichtian.

Keywords: Contourites; Palaeoceanography; Late Cretaceous; Caravaca; Betics; SE Spain

1. Introduction

Over the last few years, there have been several developments in discriminating depositional facies of contourites and their mechanisms of formation based on studies of modern deep-sea deposits (e.g., Stow et al., 1996; Stow and Faugères, 1993, 1998; Stow and Mayall, 2000a; Shanmugam, 2000). These papers, which are mainly the result of extensive deep ocean coring and seismic programmes, show that contourites are less rare in the modern record than previously thought, and also suggest that they are relatively abundant in ancient pelagic and hemipelagic sequences. Notwithstanding, only a few works have dealt in detail with ancient contourite deposits. Difficulty associated with their recognition has probably been the main cause of this oversight. However, economic

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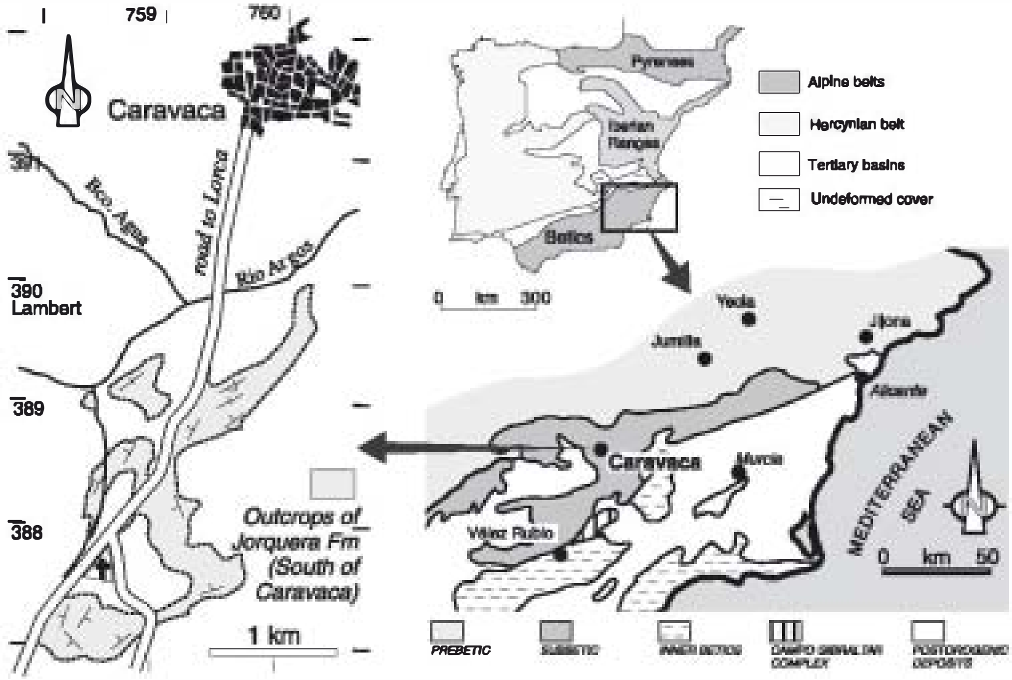

Fig. 1. Outcrops of the Jorquera Formation examined in the Caravaca area (according to Van Veen, 1969) corresponding to the mid-Campanian to mid-Maastrichtian sequence of the Subbetic zone of the Betic Chain (SE Spain).

importance of contourites as potential hydrocarbon reservoirs and their scientific relevance as palaeogeographic and palaeoceanographic indicators have recently triggered a growing interest in these particular deposits.

The aim of this paper is to provide sedimentological characteristics of carbonate contourites and document their presence on the southern palaeomargin of Iberia. The Cretaceous contourite unit analysed is of calcareous composition and crops out in the Betic chain (SE Spain, Fig. 1). We attempt to model the genesis of such deposits and their associated facies, considering changes in accommodation and accumulation. Also discussed are palaeogeographic implications of these Cretaceous contourites in a framework of the western Tethys, between Iberia and Africa.

2. Regional framework

The examined rocks crop out in the Loma de la Solana hills, about 3 km south of the town of Caravaca (Murcia province, SE Spain) (Fig. 1). This site is well known for its *K/T* boundary section. These rocks were deposited in the relatively deep, hemipelagic to pelagic settings of the ancient southern continental margin of the Iberian plate (Fig. 2). This margin was constructed during the Jurassic and Early Cretaceous as a consequence of the extensional/transtensional separation of Africa and Iberia (e.g., Savostin et al., 1986; Vera, 1988; Martín-Chivelet et al., 2002).

At the end of the Cretaceous, the tectonic evolution of the basin changed due to the convergence of Europe and Africa. As result, the broad passive margin changed into convergent (Martín-Chivelet, 1996; Martín-Chivelet et al., 1997; Reicherter and Pletsch, 2000; Chacón and Martín-Chivelet, 2001). This process culminated in the Tertiary with the complete destruction of the margin and the formation of the Betic fold-and-thrust belt (e.g., Vera, 1988). At present, the studied sequences appear as isolated outcrops within the Subbetic Zone, a major allochthonous unit of the Betics.

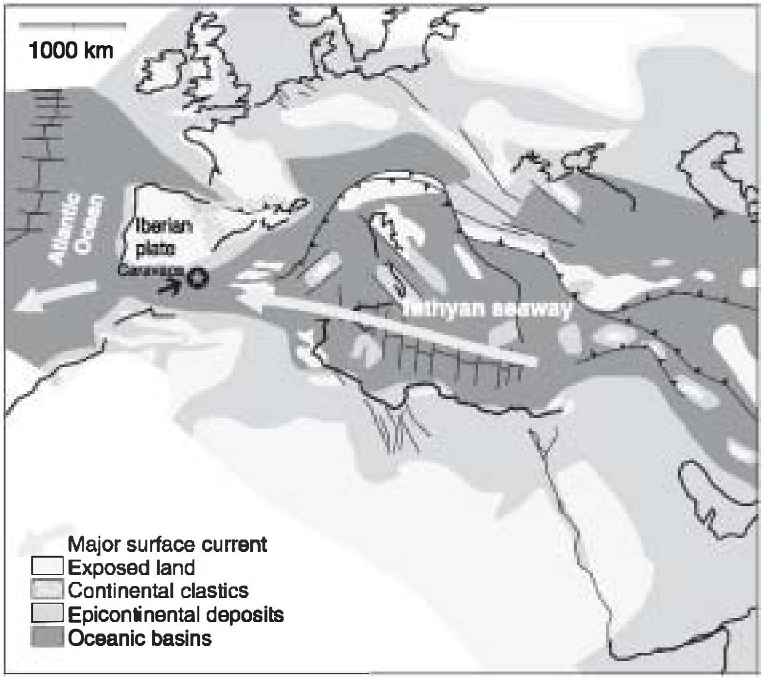

Fig. 2. Palaeogeographic reconstruction of the Tethys Seaway for the Maastrichtian (based on Philip and Floquet, 2000) showing the Betic continental margin (and the approximate position of the Caravaca outcrops) at the northern border of the gateway of the Tethys Seaway to the Central Atlantic.

In this setting close to the Iberian microcontinent were deposited (e.g., Vera and Molina, 1999) very homogeneous series of hemipelagic/pelagic carbonates during most of the Late Cretaceous, while in the shallow parts of the margin, wide carbonate (and mixed carbonate–siliciclastic) platforms developed (e.g., Azema et al., 1979; Martín-Chivelet, 1992). Deposition on both platform and deeper environments were affected by regional major tectonic events that took place in response to changes in intraplate stresses (Martín-Chivelet, 1996). These tectonic changes had short time duration and led to changes in palaeogeography, subsidence, fossils and type of sediments deposited (Martín-Chivelet, 1995).

The time span of the analysed sedimentary sequences (Late Campanian–Early Maastrichtian) corresponds to a period of relative tectonic quiescence at the margin, which lasted nearly all the Campanian and the Early Maastrichtian (i.e., about 13 million years according to Gradstein et al., 1995). It is bounded by two regional tectonic events (latest Santonian–earliest Campanian and "mid" Maastrichtian in age, respectively), which resulted in the development of two regional unconformities and their correlative conformities (Martín-Chivelet, 1995; Reicherter and Pletsch, 2000; Chacón and Martín-Chivelet, 2001).

3. Stratigraphy

Despite to the vast number of papers that refer to the *K/T* boundary in Caravaca (e.g., De Paolo et al., 1983; Bohor et al., 1986; Canudo et al., 1991; Robin et al., 1991; Kaiho and Lamolda, 1999; Arz et al., 2000), only a few general stratigraphic studies of regional scope have considered Campanian and Lower Maastrichtian deposits (e.g., Paquet, 1969; Van Veen, 1969; Hoedemaeeker, 1973; Nieto, 1997).

The sequence at Loma de la Solana Hills is a 100-m-thick succession of white, fine grained limestones and marly limestones, which include abundant dm-scale tabular intercalations of finely laminated calcarenites. The whole unit corresponds to member A of the Jorquera Formation in the sense of Van Veen (1969) (Fig. 2), thus being included in the middle part of the Quipar–Jorquera Formation in the sense of Vera et al. (1982).

The unit conformably rests on the hemipelagic carbonates of the Quipar Formation (sensu Van Veen, 1969) and is overlain by a distinct unit of distinctively greenish grey hemipelagic marls (Member B of the Jorquera Formation) of Late Maastrichtian to Paleocene age. Transition from the underlying unit is gradational and marked by a progressive increase in the number and thickness of intercalations of finely laminated calcarenite levels (herein interpreted as contourites). On the contrary, the contact with the overlying marly unit is much more abrupt, reflecting a major paleogeographic change caused by the regional tectonic event of mid-Maastrichtian age (e.g., Chacón and Martín-Chivelet, 2001; Chacón, 2002).

To precisely date the rocks and refine the age-dating proposed by Van Veen (1969), a biochronostratigraphic analysis, based on planktic foraminifera, was undertaken. The uppermost section of member A yielded the following indicative taxa: *Contusotruncana walfischensis* (Todd, 1970), *Gansserina gansseri* (Bolli, 1951), *Globotruncana aegyptiaca* (Nakkady, 1950), *Globotruncanita angulata* (Tilev, 1951), *Planoglobulina acervulinoides* (Egger, 1899), *Plummerita reicheli* (Bronnimann, 1952), *Rugoglobigerina hexacamerata* (Bronnimann, 1952) and *Rugoglobigerina scotti* (Bronnimann, 1952). This fossil assemblage characterises the middle part of the *G. gansseri* Zone, according to Robaszynski et al. (2000), and corresponds to the lower Maastrichtian (Hardenbol et al., 1998). The globotruncanid assemblage obtained from the lowermost levels of the member A is rather poor and precludes its assignation to any biozone. The very top of the underlying unit (Quipar Formation) yielded *Globotruncanita calcarata* (Cushman, 1927), a species that defines a biozone ranging in age from the late Middle Campanian to the early Late Campanian.

Comparable sequences with deposits of similar age and interpreted as contourites have also been found by the authors in several outcrops of surrounding areas of the Betics (e.g., Alicante–Jijona area, in the eastern Prebetic domain, Chacón, 2002). However, the contourite facies of these outcrops are not as well represented as in the Caravaca area.

4. Sedimentology

Two main, vertically alternating facies associations were identified in the member A of the Jorquera Formation (Fig. 3). The first consists of hemipelagites, and is clearly dominant throughout the vertical succession (more than 60% of the total thickness). The second association, which constitutes less than the 40% of the succession, is interpreted as contourite

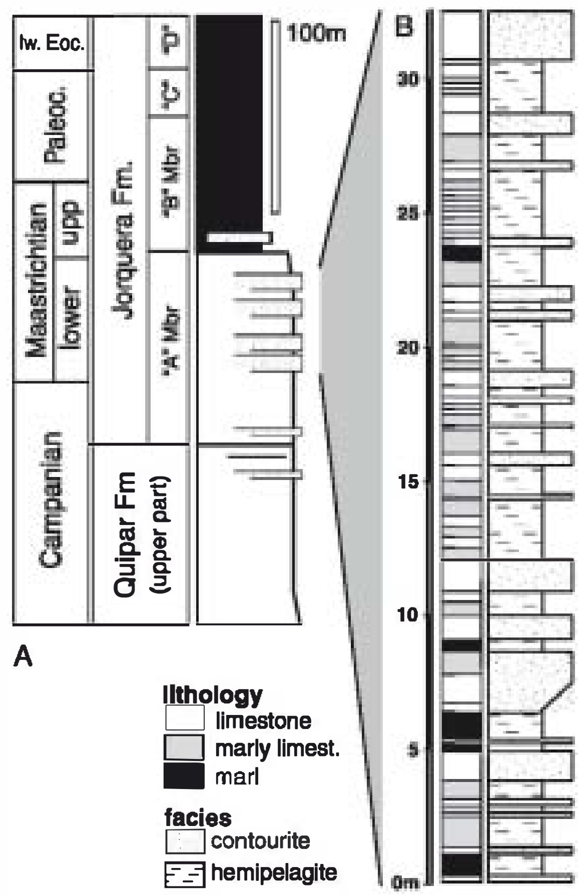

Fig. 3. (A) Latest Cretaceous–Early Palaeogene stratigraphy of the Caravaca area. Formation names and informal members were adopted from Van Veen, 1969. Ages according to Chacón (unpublished data). (B) Detailed section of the stratigraphic interval analysed where contourite and hemipelagite facies were separated.

facies, consists of sediments deposited, or significantly reworked, by deep marine currents.

4.1. Hemipelagites

The hemipelagites consist of an alternation of limestones (fine-grained wackestones) and marly limestones (fine-grained wackestones with a variable proportion of fine grained terrigenous material) (Fig. 4A), arranged in thin to medium-thick (several centimetres to a few decimetres), often poorly defined, beds. These facies characterise typical deep marine carbonate sedimentation due to the slow and relatively steady fallout of biogenic debris generated mainly in the upper part of the water column, and a variable, but always low, input of fine terrigenous material, probably transported by winds or in suspension as plumes from the land. Bioturbation is frequent but moderate, and *Chondrites* and *Zoophycos* (Fig. 4B) are the most common ichno-genera. Although *Zoophycos* and *Chondrites* have often been described in oxygen poor, organic matter-rich facies, their occurrence does not seem to imply an oxygen-depleted water/sediment interface (Bromley and Ekdale, 1984; Savrda, 1992). This point agrees with the presence of fossils of stenotopic organisms, such as inoceramids and echinoids, pointing to normal salinity and oxygenation conditions during deposition. The microfossil component includes abundant globotruncanids, calcisphaerulids and some deep-water benthic foraminifera, which appear randomly distributed and show no evidence of reworking or sorting. This suggests minor or no current influence on deposition.

The benthic foraminifera assemblage—*Gaudryina pyramidata* (Cushman, 1926); *Lenticulina* sp.; *Marssonella oxycona* (Reuss, 1860); *Oridorsalis* sp.; *Reussella szajnochae* (Grzybowski, 1896), among others—points to an upper bathyal palaeo-waterdepth, i.e., between 200 and 600 m (Van Morkhoven et al., 1986). Hemipelagic deposits from the uppermost part of the underlying Quipar Formation have yielded a richer benthic assemblage with *G. pyramidata* (Cushman, 1926), *Gavelinella becariiformis* (White, 1928), *Lenticulina* sp., *Oridorsalis* sp., *Quadrimorphina* sp., *R. szajnochae* (Grzybowski, 1896) and *Stensioina pommerana* (Brotzen, 1940), which confirms this bathymetric interpretation.

The presence of alternating decimetre-scale limestones and marly limestones suggests a high-frequency cyclicity recorded throughout the hemipelagites that is not observed where abundant contourite deposits occur (Fig. 4A). This type of cyclicity is a pervasive feature of similar hemipelagic facies in the

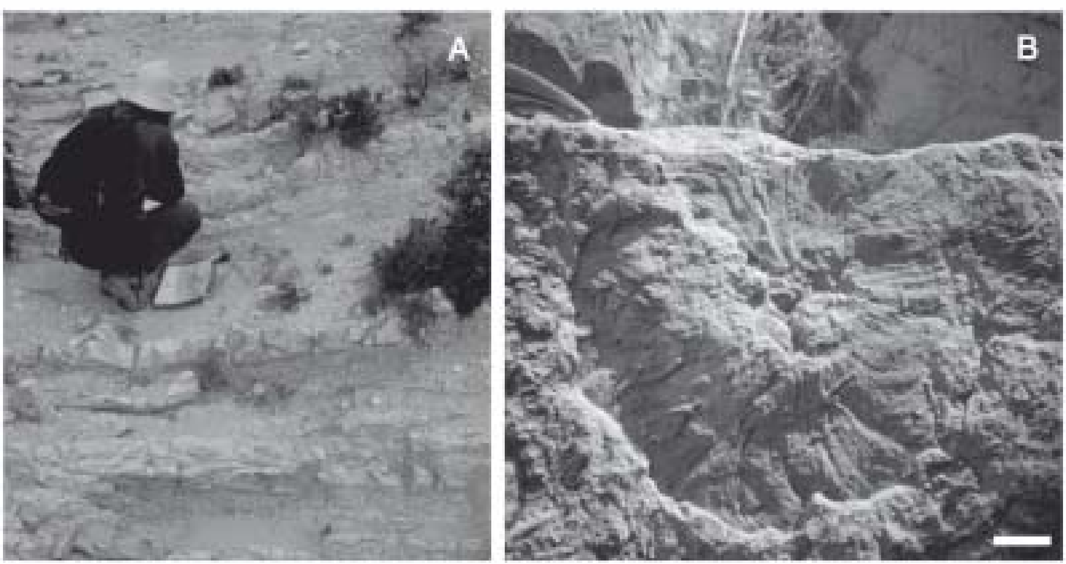

Fig. 4. Hemipelagite facies. (A) Typical rhythmic arrangement of limestones and marly limestones on the outcrop scale. (B) Detail of a *Zoophycos* structure within these facies (scale bar: 5 cm).

underlying Quipar Formation, as well as in coeval sediments in other areas in the basin. It has been interpreted to be related to climate changes controlled by Milankovitch orbital cycles (e.g., Vera and Molina, 1999; Chacón and Martín-Chivelet, 2001).

4.2. Contourites

Contourites appear as stratified, tabular, vertically isolated beds 0.05–1.6 m thick (Fig. 5A). The base and top of each contourite bed are sharp, and mark the change from, or into, the hemipelagic facies. The base is often markedly erosive and can show scour cast structures. Flute casts are a few millimetres deep and isolated U-shaped longitudinal scour casts resemble gutter cast morphologies, but are smaller than the usual gutter structures. Flute casts are massive whereas longitudinal scour marks are filled with cross-laminated calcarenites. In contrast, the tops are never erosive and are usually defined by a flat, ripple-free surface. Sometimes, two or more contourite beds can be amalgamated, and are thus separated by erosive surfaces.

Contourites are formed by very fine to medium sand-grained calcarenites which can occasionally reach a larger grain size. Texturally, they are carbonate grainstones to packstones, with usually less than 5% of siliciclastic grains of silt to fine sand size. Bioclastic material is dominant and consists of highly reworked and fragmented tests of the same organisms that appear in the hemipelagites (planktic foraminifera, a few deep water benthic foraminifera and calcisphaerulids, as well as calcite prisms arising from the break up of inoceramid shells). The bioclasts are mixed with abundant subrounded to angular intraclasts of micritic material similar to that predominating in the hemipelagic facies.

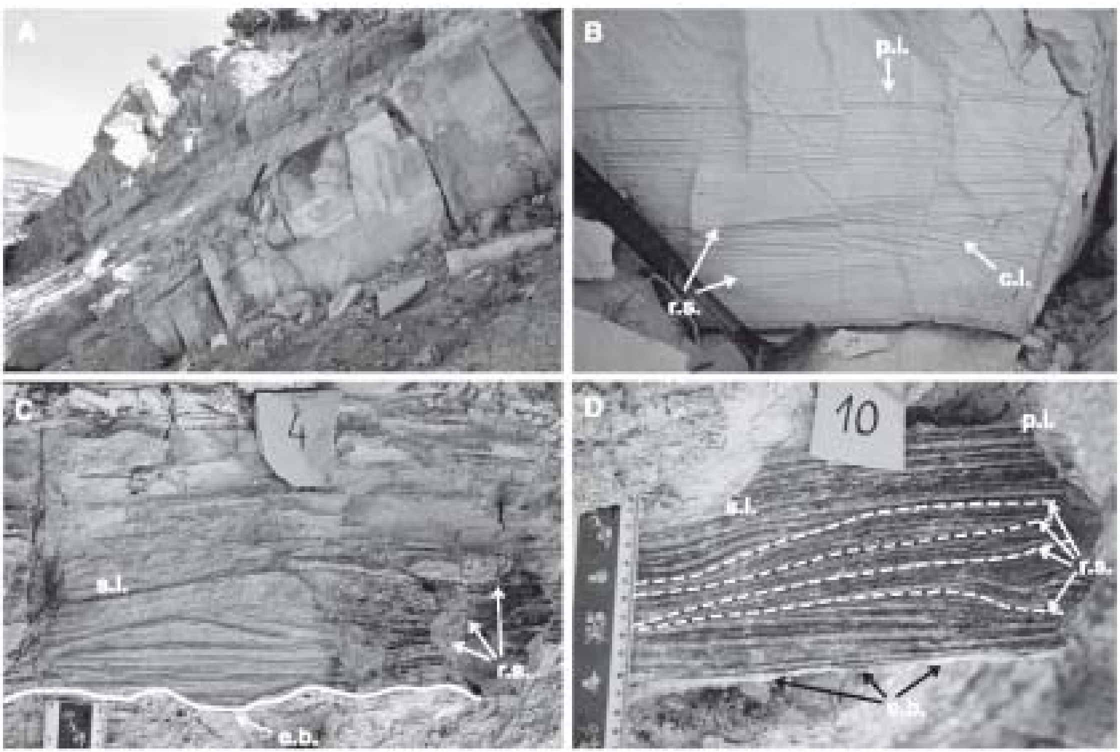

Fig. 5. Examples of contourite facies in the "A" member of the Jorquera Formation (Lower Maastrichtian). (A) Outcrop feature of dm to m scale contourite beds interbedded with less competent hemipelagic facies. See hammer for scale. (B to D) Details of contourite facies showing the most typical sedimentary structures and their spatial distribution (e.b.: erosive basal surface; r.s.: internal reactivation surfaces, usually involving erosion; c.l.: cross lamination; s.l.: sinusoidal lamination; p.l.: parallel lamination). Scale in cm.

Contourite beds are characterised by a defined vertical distribution sequence of both sedimentary structures and grain size (Fig. 6). Vertical grain size variations are observed at different scales. Both normal and reverse gradings are present within each contourite bed, although positive grading predominates. Maximum grain sizes are commonly found in the lower half of each level, but not necessarily at the base. Thereafter, it is common to observe an increase in grain size from the base to a certain level located a few centimetres above it, and a slight and gradual decrease from this level to the top of the bed. According to the observed sedimentary structures, most contourite beds can be roughly divided into two distinct intervals:

(1) A lower interval, which may represent a third to as much as over half the entire bed, characterised by the dominance of cross lamination, produced by the migration of ripples under the influence of a tractive current. Larger bedforms are only occasionally observed. Reactivation and slightly erosive surfaces are also frequent in this interval. Within this lower part, a vertical sequence of structures is often observed: above the basal erosive surface of the contourite bed, the first, overlying deposit usually consists of a thin level (5–20 cm, Fig. 5B) with parallel to low angle cross lamination which often shows reverse grading. This level is covered by a new deposit, in which sigmoidal, cross lamination is the most prominent feature (Fig. 5C–D). This sigmoidal lamination is the result of the migration of current ripples, mainly climbing ripples, although non-climbing ones can also be found. Most sets have been partially eroded by reactivation, but when observed to be complete, they are strongly asymmetrical, with depositional stoss sides showing extremely gentle slopes (1–4°) and steeper lee sides (10–15°). The height of the ripples never exceeds 2 or 3 cm, and their wavelength averages 10 cm. The angle of ripple climbing is low, about 20° or less. Foreset laminae are typically concave upwards and 1–2 mm thick. Mud-offshots are very common. These ripples can be ascribed to the type B of Allen (1970, 1971) and Ashley et al. (1982), type III of Yokokawa et al. (1995) and, according to the nomenclature of Hunter (1977), they are supercritical. The number of ripples in each train is low, compared to the climbing ripples usually described in turbidites and continental facies. Preservation of thick cosets of cross lamination is rather rare, since internal reactivation surfaces are frequent (Figs. 5B to D, 7 and 8). These surfaces may correspond to erosive surfaces that involve truncation of the previous bedforms; though true and deep erosion is sometimes absent or very slight, and the observed "unconformities" are due to the geometry resulting from prominent changes in the migration direction of ripple trains. Such variations are the norm from one episode of migration to the next. When it is not possible to observe structures in all three dimensions, reactivation surfaces may be masked and a subtle change in the angle of inclination of the lee and stoss sides or in the angle of climbing is the only evidence of discontinuity in current action (Fig. 7). When there are signs of ripple evolution, climbing ripples are found to merge laterally and pass upwards to an interval of sinusoidal lamination (sensu Jopling and Walker, 1968) or draped lamination (sensu Gustavson et al., 1975; Ashley et al., 1982). This sinusoidal lamination characterises the transition between the two parts into which the contourite bed has been divided.

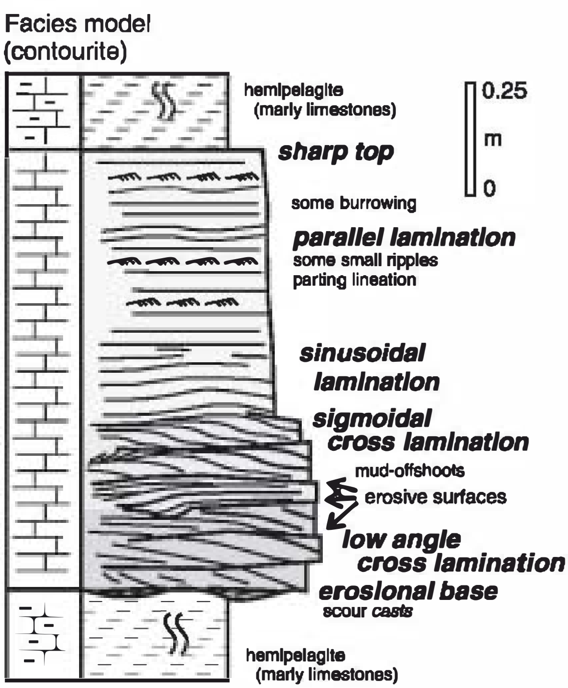

Fig. 6. General facies model proposed for the contourites of Caravaca.

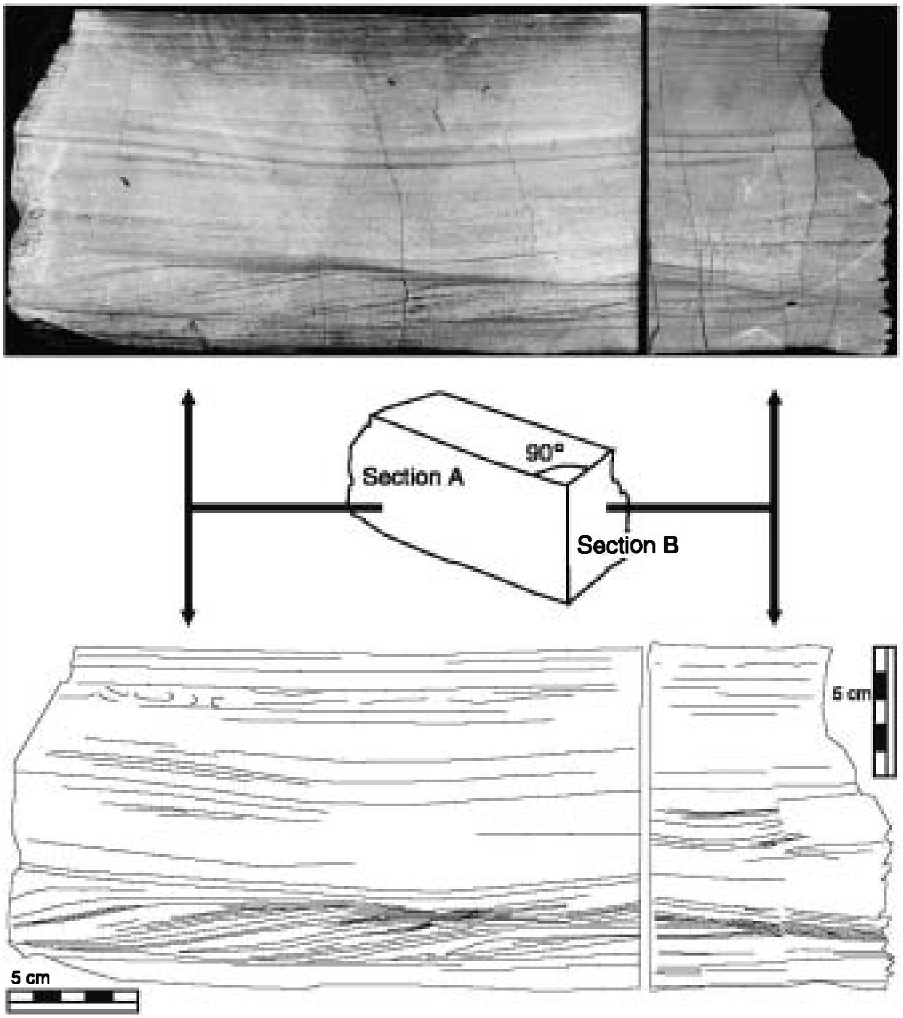

Fig. 7. Two polished sections of a contourite bed, where some of the main sedimentary structures of contourites described in the text are shown: sigmoidal cross-lamination, draped lamination, parallel lamination and reactivation surfaces. Note the change in dipping of foresets located under and above the reactivation erosive surface clearly reflecting a change in the direction of ripple migration and thus in current direction.

(2) An upper interval, which is dominated by very thin and parallel lamination, grain size being generally finer than the underlying interval. The upper part of the contourites is mainly defined. These structures arise from changes in grain size and micrite content, and are accentuated by concentrations of small foraminifera and inoceramid shell fragments. Thin beds are of micrometric to millimetric order and often show sharp bases, in which parting alignment can be observed. Occasionally, normal grading can be observed in the laminae, which tend to cross at extremely low angles. Very small lenses of calcarenite or silt, which probably correspond to starved ripples, appear in some micritic laminae.

The sequences of the structures described above lead us to conclude that the entire bed was deposited by contourite currents. Such strata represent the deposit of a multi-episodic, long-standing, bottom tractive current which decreases in strength and average velocity over time and in space as its ratio of suspended load/bed load increases. On a smaller scale, the internal distribution of structures in the lower part of the bed also indicates the action of de-accelerating, tractive currents that were repeatedly reactivated.

Fig. 8. Cross lamination at the microscopic scale in a contourite bed. Scale bar: 1 mm.

The onset of current action on the sea bottom is marked by erosion at the base of each bed. After that, the current de-accelerates and contourite facies sedimentation begins. During this first stage of deposition, variations in current strength occur, causing minor erosive surfaces as well as vertical changes in grain size. Both reverse and normal grading have been recognised in this basal interval.

The development of climbing ripples and the associated abundance of mud-offshots are indicative of bottom low density traction currents with considerable amounts of suspended load. Several scales of evolution, from type B ripple-drift cross lamination to draped or sinusoidal lamination, probably involved an increase in suspended load and/or a decrease in current velocity. In this situation, the net rate of aggradation is high and tends to surpass the downstream migration rate of ripples.

Changes in the characteristics of the sedimentary structures and the abundance of reactivation surfaces indicate that current strength is variable and its influence on the sea floor discontinuous. Parameters defining the current, such as velocity, direction or the ratio of suspended load/bed load, repeatedly changed as the deposits were generated. A first estimation of current velocity can be made by comparing our observations with those emerging from experimental studies by Ashley et al. (1982). According to work performed on medium-sized sand, type B ripples required velocities of 15–25 cm/s to be formed, whereas draped lamination needs less than 15 cm/s. Since our deposits are very fine-grained calcarenites, lower velocities, probably in the range 10–15 cm/s, would be expected to be sufficient for the formation of the structures observed.

The upper part of the contourite beds is interpreted to be the result of aggradation and sedimentation processes due to currents that mainly transport suspended load, with episodes of activation of extremely weak tractive currents.

Environmental dynamics inferred from the sedimentological analysis of contourites and hemipelagites and the stacking pattern shown by these facies (Fig. 3) suggest that the entire stratigraphic succession formed in the hemipelagic realm of the basin at the upper bathyal palaeo-waterdepth, i.e., between 200 and 600 m., within a palaeogeographic domain that was periodically affected by currents that actively modified the bottom. These currents were able to transport and deposit fine sand to silt sized hemipelagic sediments.

Current strength and the effects of its action on the bottom changed throughout the deposition of every bed of contourites and hemipelagites. The sequence in Fig. 6, representing one episode of contourite deposition, can be explained by a succession of four main situations, each characterised by the following environmental and sedimentary conditions:

1. A strongly erosive current and the development of erosive surfaces develop but no net sedimentation.
2. Active influence of currents of variable intensity and periods of erosion alternating with periods of deposition. Contourite facies resulting from this situation show internal erosive surfaces separating deposits of cross-laminated calcarenites.
3. Decreasing effects of currents that are active but of low strength. The material transported by currents is deposited and the resulting facies are dominated by fine-grained, horizontally laminated contourites that rarely show bioturbation.

4. Very weak or no influence of currents on the sea floor results in "normal" hemipelagic sedimentation.

5. Latest Cretaceous ocean circulation

The genesis of mid-depth contourite deposits requires the presence of either relatively strong, semi-permanent geostrophic currents flowing at intermediate depths or very strong superficial or subsuperficial currents able to affect the sea floor at these depths (Viana et al., 1998). It is likely that one of these current types controlled the deposition of the latest Cretaceous calcareous contourites of the Subbetic. In ancient contourites, however, it is difficult to discern among the different types of currents that might have controlled their deposition. In the present study, this interpretation is even more difficult, since the sequences considered were strongly tectonically deformed, impeding any reliable palaeocurrent analysis. For these reasons, we adopt a more speculative approach and consider whether the palaeoceanographic and palaeogeographic conditions of the Betic basin could have been appropriate for the generation of such currents.

Knowledge on ocean circulation during Late Cretaceous times is very limited. Palaeobiogeographical reconstructions and computer simulations of ocean currents during the Late Cretaceous suggest that a circumglobal tropical surface current flowed westwardly through the continental configuration of that time (e.g., Haq, 1984; Bush, 1997). This current was determined by the opening, during the Mesozoic, of wide and deep latitudinal ocean passages in the intertropical area as a consequence of the break-up of Pangea and the divergence of Eurasia and Africa (Tethys Seaway) and of North and South America (Deep Central American Seaway). The latter was induced by the easterly wind stress that dominated (as does today) in latitudes below 30°. These easterlies, in conjunction with the open ocean gateways between the Indian, Tethys, and Pacific ocean basins, drove the circumglobal, westward-flowing current, which flowed along the equator in the Pacific Ocean, around the southern end of the Indonesian peninsula, into the Tethys Seaway between Africa and Eurasia and finally back into the Pacific Ocean through the ocean gateway between North and South America (Fig. 2).

Through general circulation modelling (GCM), Bush (1997) presented a numerical simulation for the Late Cretaceous tropical oceans of higher resolution than the former ones (e.g., Barron and Peterson, 1990), in which the tropical circumglobal current appears as a robust feature of ocean circulation for that time. Bush used a coupled atmosphere–ocean model and adopted a palaeogeographic reconstruction for the Maastrichtian. According to this simulation, the mean depth of the circumglobal tropical current varies substantially but attained its vertical maximum in the Tethys Seaway, where it reached 350 m. It is possible that, below that strong surface current, a complex pattern of eastward subsurface currents (or undercurrents) developed in the Tethys Seaway to compensate that immense westward water surface flow. Both the main surface current system and the associated undercurrents could be responsible of contourite deposition in the South Iberian continental margin.

During the Late Cretaceous, the Betic continental margin occupied an area that was extremely susceptible to one of these relatively deep oceanic currents. Situated in the western end of the Tethys Seaway passage at this ocean's gateway to the Atlantic (Fig. 2), the continental margin formed the northern boundary of the narrowest oceanic passage through which the broad equatorial current system flowed. Relatively deep currents, channelled between Africa and Iberia, could have affected the ocean floor in the hemipelagic areas of the continental margin and generated contourite sequences.

At this point, a question arises: why this process was effective in the Campanian–Maastrichtian interval and not before, despite that the Tethyan passage was effective during nearly all of the Mesozoic?

Possible explanations to this time restriction should be found in the palaeogeographic evolution that took place at the end of the Cretaceous. The whole intertropical current system was strongly sensitive to palaeogeographic evolution, representing sea-level changes or tectonic plate movement (e.g., Poulsen et al., 1998). Late Cretaceous ocean circulation in the Tethys was substantially different from the previous Early Cretaceous–Late Jurassic ocean circulation, the main differences being related to the start of the collision of India with Asia and the onset of Africa–Europe convergence.

In particular, this later point could cause a paleoceanographic rearrangement in the Tethyan passage, where the Caravaca contourites are deposited. Small changes in palaeogeography could provoke drastic modifications in the direction, strength and depth of the currents, and thus condition their effects in hemipelagic settings. From Late Santonian times onwards, Alpine convergence started to seriously affect the sedimentary basins of Iberia (Pyrenean basin, Iberian basin, Betic margin), leading to the onset of basin inversion in the Pyrenees (e.g., Simó, 1986) and the Iberian basin (e.g., Alonso et al., 1993), and to strong tectonic movements in the Betics (e.g., Martín-Chivelet, 1996; Reicherter and Pletsch, 2000). Narrowing of the passage between Iberia and Africa with the consequent intensification of channelling water transfer between the Tethys and the Atlantic was also possibly induced. Thus, the broad geodynamic process related to the onset of alpine convergence caused palaeogeographic changes in the Betic margin, including block movements and regional uplift (Reicherter and Pletsch, 2000). This led to pronounced changes in the configuration of the basin floor, and as a consequence, to intensification and deepening of oceanic currents in the area, favouring contourite generation.

6. Discussion and conclusions

Since Wüst (1936, 1958) suggested that bottom currents might have a prominent role in reworking and distributing deep ocean sediments, many authors have underscored the importance of sediment reworking processes by bottom currents (Kelling, 1958, 1964; Heezen, 1959; Craig and Walton, 1962; Kuenen, 1964, 1967; Hsü, 1964; Ballance, 1964; Heezen and Hollister, 1964; Dzulynski and Walton, 1965; Bouma, 1972, 1973). The earliest identifications of deposits clearly ascribable to contour currents and consequent definition of contourites were undertaken by Heezen et al. (1966) and Wezel (1969), and documented by Hollister (1967), Hollister and Heezen (1967, 1972), Bouma and Hollister (1973), Hollister et al., (1974a,b), Damuth (1975) and Stow (1977) among others.

The DSDP, and especially the HEBBLE programme, developed during the late 1970s and early 1980s has greatly improved our understanding of bottom currents and associated sediments (Hollister and McCave, 1984; McCave and Hollister, 1985; Nowell and Hollister, 1985). In contrast, the definitions and criteria to recognise contourites were established by Stow (1977, 1982), Stow and Bowen (1978), Stow and Lovell (1979), Lovell and Stow (1981), Stow et al. (1986) and by the papers compiled in a special issue edited by Stow and Piper (1984a). Many of these papers deal with the problem of distinguishing fine-grained turbidites from contourites. This issue was further explored in published facies models for fine-grained turbidites (Stow and Shanmugam, 1980; Stow and Piper, 1984b; Piper and Stow, 1991). Subsequent facies models for contourites (Stow et al., 1996, 1998), studies on the contourite–turbidite continuum (Stanley, 1993) and later revisions of previous cases in light of the new models and the criteria provided by them (Faugères and Stow, 1993; Stow et al., 1998; Stow and Mayall, 2000b) may have served to resolve the controversy that began 40 years ago. However, authors such as Shanmugam et al. (1993, 1995) and Shanmugam (2000) propose alternative criteria to those of Faugères and Stow (1993) and Stow et al. (1996) for defining and recognising contourites.

After careful revision of the literature and comparison of the sediments described with the criteria followed by the different authors in the study of recent and ancient deposits, it becomes clear that there are still many unresolved questions related to contourites, and there is a general lack of consensus concerning what can be exactly ascribed to deep current deposits. The main conclusion to be drawn from this revision is that contourites or "sediments in relative deep water, deposited or significantly reworked by stable geostrophic currents (Heezen et al., 1966; Faugères and Stow, 1993)" present high variability in both recent and ancient records (e.g., papers compiled in Stow and Faugères, 1993, 1998; Stow and Mayall, 2000a). This inherent variability depends on factors such as water depth, current intensity, type and supply of available sediment and biological activity. This variability and perhaps the lack of a true understanding of the similarities and differences between modern and ancient bottom currents (Shanmugam, 2000) may explain why the construction of a general facies model for these deposits is proving to be such an arduous task, and why the already available models, based on recent deposits, are of limited value and application to the stratigraphic record.

The calcarenitic deposits of the Late Cretaceous of Caravaca examined here do not entirely match any of the published models, although they do share some features with them. Their interpretation, however, as contourites or deposits of bottom currents, is the most plausible hypothesis that can be proposed after the careful sedimentological, stratigraphic and palaeogeographic analysis performed. This hypothesis is supported by the following features.

6.1. Composition and texture of sediments

As previously described, the carbonate contourites of Caravaca consist of very fine to medium-grained calcarenites mainly composed of variable yet rather small amounts of siliciclastic grains of silt to fine sand-sized and bioclastic material, highly fragmented planktic foraminifera, a few deep water benthic foraminifera and calcisphaerulids. There is also a presence of calcite prisms from the break up of inoceramid tests, and abundant subrounded to angular intraclasts, made up of micritic material similar to that predominating in the hemipelagic facies. That is, sedimentary particles from the hemipelagic realm that have undergone transport within the same environment as where they were produced. They do not contrast in composition with the interbedded hemipelagites, as commonly do fine-grained or mud turbidites (Piper and Stow, 1991). The association of foraminifers that characterise the contourites of Caravaca is completely different to any described in contemporaneous and laterally equivalent deposits of shallower areas of the basin (Martín-Chivelet, 1995). In younger units, also formed in the hemipelagic domain of the basin, turbidite facies become abundant and show, at least, a certain mix of shallow and deep-water faunas.

6.2. Structures and inferred current behaviour

Sediments are clean and well sorted, and the sedimentary structures observed point to traction as the main mechanism of transport. Since traction is not uncommon in turbidity currents, this evidence does not preclude such a mechanism of sedimentation acting by itself, yet no evidence of transport by gravity-driven currents has been found. Furthermore, multiple reactivation surfaces within each calcarenite bed reveal alternating episodes of erosion and sedimentation, which are not characteristic of turbidites. As described in this paper, the contourites of Caravaca were controlled by multi-episodic, long-standing, tractive bottom currents of variable current strength and direction. Most of the structures observed were probably formed under current speeds in the range of 10–15 cm/s. These currents are comparable in many aspects to those described for the present-day oceans by Hollister et al. (1984), Hollister and McCave (1984) and Hollister (1993), defined as "benthic storms".

6.3. Palaeogeographic context

The sedimentary record of the Betic margin in the Late Cretaceous and Early Tertiary reflects palaeogeographic homogenisation of the Early Cretaceous complex basin configuration (e.g., Vera, 1988) and later the start of the transition of the passive margin into a convergent one (Martín-Chivelet, 1996; Reicherter and Pletsch, 2000). During the Campanian and Early Maastrichtian, this basin homogenisation was still patent, with highly uniform hemipelagic to pelagic sedimentary conditions elsewhere in the deep settings of the margin and without the development of turbidite sequences or other gravitational deposits (e.g., Vera and Molina, 1999; Chacón and Martín-Chivelet, 2001). The platform to basin transition for this interval is defined by a depositional, very flat outer ramp, in which gravitational structures such as slumps are scarce (Martín-Chivelet, 1995). However, overlapping this homogenisation, a process of regional uplift, which probably mirrors the start of convergence, took place. This process probably created the necessary bathymetric conditions for the contourite sequences to be generated.

6.4. Palaeohydrological context

During the Late Cretaceous, the Betic margin occupied a palaeogeographic position that was susceptible to the effects of relatively deep oceanic currents related to the circumglobal equatorial current system. This led to the formation of the northern margin of the narrowest oceanic passage, through which the broad equatorial current system flowed:

the Tethys Seaway gateway into the Central Atlantic. Oceanic currents, channelled between Africa and Iberia, are sure to have affected the sea bottom in the hemipelagic areas of the continental margin.

In conclusion, this paper describes a case study of calcareous contourite facies, of Late Campanian to Early Maastrichtian age, that were deposited in the almost flat, wide areas of mid-depth settings (200–600 m) of the Mesozoic Betic continental margin. In this palaeogeographic domain, combined tectonic and palaeoceanographic conditions towards the end of the Cretaceous led to the generation of an alternating succession of contourite and hemipelagite beds. The contourite beds are characterised by a sequence of traction structures, internal erosive surfaces, inverse and normal grading and other features which allow us to propose a new facies model. This model is compared to those described in literature, and attempts to contribute towards a better understanding of sedimentary processes in ancient deep oceans. Our intention was not to provide an alternative model to those already published, but to complement these by supplying new data derived from a context that is fairly different to that of the present-day oceans. The main differences in our model arise from the nature and composition of the sediments involved and the patterns of latest Cretaceous oceanic circulation, extensively controlled by global palaeogeographic constraints due to the different distribution of continents. Neither of these conditions have been considered in the currently accepted facies models, probably since we are presently unaware of any similar situation. Nonetheless, the combination of both these conditions may serve to explain differences in the facies themselves, in their arrangement and in stratigraphic stacking patterns.

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References

Alonso, A., Floquet, M., Mas, R., Meléndez, A., 1993. Late Cretaceous Carbonate Platforms: Origin and Evolution, Iberian Ranges, Spain. In: Simo, J.A.T., Scott, R.W., Masse, J.P. (Eds.), Cretaceous Carbonate Platforms. AAPG Memoir, vol. 56, pp. 297–314.

Allen, J.R.L., 1970. A quantitative model of climbing ripples and their cross-laminated deposits. Sedimentology 14, 5–26.

Allen, J.R.L., 1971. Instantaneous sediment deposition rates deduced from climbing-ripple cross-lamination. J. Geol. Soc. 127, 553–561.

Arz, J.A., Arenillas, I., Molina, E., Sepúlveda, R., 2000. La estabilidad evolutiva de los foraminíferos planctónicos en el Maastrichtiense superior y su extinción en el limite Cretácico/Terciario de Caravaca, España. Revista Geológica de Chile 27, 27–47.

Ashley, G.M., Southard, J.B., Boothroyd, J.C., 1982. Deposition of climbing-ripple beds: a flume simulation. Sedimentology 29 (1), 67–79.

Azema, J., Foucault, A., Fourcade, E., García-Hernández, M., González-Donoso, J.M., Linares, A., Linares, D., López-Garrido, A.C., Rivas, P., Vera, J.A., 1979. Las microfacies del Jurásico y Cretácico de las Zonas Externas de las Cordilleras Béticas. Secretariado de Publicaciones de la Universidad de Granada, Granada. 83 pp.

Ballance, P.F., 1964. Streaked out mud ripples below Miocene turbidites, Puriri Formation, New Zealand. J. Sediment. Petrol. 34, 91–101.

Barron, E.J., Peterson, W.H., 1990. Mid-Cretaceous Ocean Circulation: Results from Model Sensitivity Studies. Paleoceanography 5 (3), 319–337.

Bohor, B.F., Foor, E.E., Ganapathy, R., 1986. Magnesioferrite from the Cretaceous-Tertiary boundary, Caravaca, Spain. Earth Planet. Sci. Lett. 81, 57–66.

Bouma, A.H., 1972. Recent and ancient turbidites and contourites. Trans. - Gulf Coast Assoc. Geol. Soc. 22, 205–221.

Bouma, A.H., 1973. Contourites in Niessenflysch, Switzerland. Eclogae Geol. Helv. 66 (2), 315–323.

Bouma, A.H., Hollister, C.D., 1973. Deep Ocean Basin Sedimentation. In: Middleton, G.V., Bouma, A.H. (Eds.), Turbidites and Deep Water Sedimentation. Soc. Econ. Paleontol. Mineral., Tulsa, Okla, pp. 79–118.

Bromley, R.G., Ekdale, A.A., 1984. *Chondrites*: a trace fossil indicator of anoxia in sediments. Science 224, 872–874.

Bush, A.B.G., 1997. Numerical Simulation of the Cretaceous Tethys Circumglobal Current. Science 275, 807–810.

Canudo, J.I., Keller, G., Molina, E., 1991. Cretaceous-Tertiary boundary extinction pattern and faunal turnover at Agost and Caravaca, SE Spain. Mar. Micropaleontol. 17, 319–341.

Chacón, B., 2002. Las sucesiones hemipelágicas del final del Cretácico e inicio del Paleógeno en el SE de la Placa Ibérica: Estratigrafía de eventos y evolución de la cuenca. Unpubl. PhD Thesis. Universidad Complutense, Madrid. 439 pp.

Chacón, B., Martín-Chivelet, J., 2001. Implicaciones tectosedimentarias de la discontinuidad estratigráfica del Maastrichtiense medio en Aspe (Prebético de Alicante). Rev. Soc. Geol. Esp. 14 (1–2), 123–133.

Craig, G., Walton, K., 1962. Sedimentary structures and paleocurrents directions from the Paleozoic rocks of Kircudbrightshire. Trans. Edinb. Geol. Soc. 19, 100–119.

Damuth, J.E., 1975. Echo-character of the western equatorial Atlantic floor and its relationship to the dispersal and distribution of terrigenous sediments. Mar. Geol. 18, 17–45.

De Paolo, D.J., Kyte, F.T., Marshall, B.D., O'Neil, J.R., Smith, J., 1983. Rb-Sr, Sm-Nd, K-Ca, O and H isotopic study of Cretaceous-Tertiary boundary sediments, Caravaca, Spain: Evidence for an oceanic impact site. Earth Planet. Sci. Lett. 64, 356–373.

Dzulynski, S., Walton, E.K., 1965. Sedimentary features of flysch and graywackes. Elsevier, Amsterdam. 274 pp.

Faugères, J.C., Stow, D.A.V., 1993. Bottom-current-controlled sedimentation: a synthesis of the contourite problem. In: Stow, D.A.V., Faugères, J.C. (Eds.), Contourite and Bottom Currents. Sediment. Geol., vol. 82, pp. 287–297.

Gradstein, F.M., Agterberg, F.P., Ogg, J.G., Hardenbol, J., Van Veen, P., Thierry, J., Huang, Z., 1995. A Triassic, Jurassic and Cretaceous Time Scale. In: Berggren, W.A., Kent, D.V., Aubry, M.P., Hardenbol, J. (Eds.), Geochronology, Times Scales and Global Stratigraphic Correlation, vol. 54. Society of Economic Paleontologists and Mineralogists Special Publication, Tulsa, pp. 95–126.

Gustavson, T.C., Ashley, G.M., Boothroyd, J.C., 1975. Depositional sequences in glaciolacustrine deltas. In: McDonald, B.C., Jopling, A.V. (Eds.), Glaciolacustrine sedimentation. Soc. Econ. Paleont, Miner., vol. 3, pp. 264–280. Tulsa.

Haq, B.U., 1984. Palaeoceanography: A synoptic overview of 200 million years of ocean history. In: Haq, B.U., Milliman, J.D. (Eds.), Marine Geology and Oceanography of Arabian Sea and Coastal Pakistan. Van Nostrand Reinhold, New York, pp. 201–231.

Hardenbol, J., Thierry, J., Farley, M.B., Jacquin, Th., Graciansky, P.C., de Vail, P.R., 1998. Mesozoic and Cenozoic sequence stratigraphic framework of European basins. Mesozoic and Cenozoic Sequence Stratigraphy of European basins, vol. 60. SEPM Special Publication, Tulsa, OK.

Heezen, B.C., 1959. A review of natural processes contributing to the mechanical failure of deep-sea cables. Geophys. J. 2, 142–163.

Heezen, B.C., Hollister, C.D., 1964. Deep-sea current evidence from abyssal sediments. Mar. Geol. 1, 141–174.

Heezen, B.C., Hollister, C.D., Ruddiman, W.F., 1966. Shaping of the continental rise by deep geostrophic contour currents. Science 152, 502–508.

Hoedemaeker, Ph.J., 1973. Olisthostromes and other delapsional deposits, and their occurrence in the region of Moratalla (Prov. of Murcia Spain). Scr. Geol. 19, 1–207.

Hollister, C.D., 1967. Sediment distribution and deep circulation in the Western North Atlantic. PhD Thesis, Columbia University, New York, 467 pp.

Hollister, C.D., 1993. The concept of deep-sea contourites. In: Stow, D.A.V., Faugères, J.C. (Eds.), Contourites and Bottom-Currents. Sediment. Geol., vol. 82, pp. 5–11.

Hollister, C.D., Heezen, B.C., 1967. Contour current evidence from abyssal sediments. Trans. - Am. Geophys. Union 48, 142.

Hollister, C.D., Heezen, B.C., 1972. Geologic effects of ocean bottom currents: western North Atlantic. In: Gordon, A.L. (Ed.), Studies in Physical Oceanography, vol. 2. Gordon and Breach, New York, pp. 37–66.

Hollister, C.D., McCave, I.N., 1984. Sedimentation under deep-sea storms. Nature 309 (5965), 220–225.

Hollister, C.D., Johnson, D.A., Lonsdale, P.F., 1974a. Current-controlled abyssal sedimentation: Samoan passage, equatorial west Pacific. J. Geol. 82, 275–300.

Hollister, C.D., Flood, R.D., Johnson, D.A., Lonsdale, P., Southard, J.B., 1974b. Abyssal furrows and hyperbolic echo traces on the Bahama Outer Ridge. Geology 2, 395–400.

Hollister, C.D., Nowell, A.R.M., Jumars, P.A., 1984. The dynamic abyss. Sci. Am. 250 (3), 42–53.

Hsü, K.J., 1964. Cross lamination in graded bed sequences. J. Sediment. Petrol. 34, 379–388.

Hunter, R.E., 1977. Terminology of cross-stratified sedimentary layers and climbing ripples structures. J. Sediment. Petrol. 47, 697–706.

Jopling, A.V., Walker, R.G., 1968. Morphology and origin of ripple drift cross-lamination, with examples from the Pleistocene of Massachusetts. J. Sediment. Petrol. 38, 971–984.

Kaiho, K., Lamolda, M.A., 1999. Catastrophic extinction of planktonic foraminifera at the Cretaceous-Tertiary boundary evidenced by stable isotopes and foraminiferal abundance at Caravaca, Spain. Geology 27, 355–358.

Kelling, G., 1958. Ripple-mark in the Rhinns of Galloway. Trans. Edinb. Geol. Soc. 17, 117–132.

Kelling, G., 1964. The turbidite concept in Britain. In: Bouma, A.H., Brouwer, A. (Eds.), Turbidites. Elsevier, Amsterdam, pp. 75–92.

Kuenen, Ph.H., 1964. Deep-sea sands and ancient turbidites. In: Bouma, A.H., Brouwer, A. (Eds.), Turbidites. Elsevier, Amsterdam, pp. 3–33.

Kuenen, Ph.H., 1967. Emplacement of flysch-type sand beds. Sedimentology 9, 203–243.

Lovell, J.P.B., Stow, D.A.V., 1981. Identification of ancient sandy contourites. Geology 9, 347–349.

Martín-Chivelet, J., 1992. Las plataformas carbonatadas del Cretácico superior de la Margen Bética (Altiplano de Jumilla Yecla, Murcia) Col. Tesis Doctorales, Universidad Complutense, Madrid. 899 pp.

Martín-Chivelet, J., 1995. Sequence stratigraphy of mixed carbonate–siliciclastic platforms developed in a tectonically active setting, Upper Cretaceous, Betic Continental Margin (Spain). J. Sediment. Res. b65, 235–254.

Martín-Chivelet, J., 1996. Late Cretaceous Stratigraphic Patterns and subsidence history of the Betic Continental Margin (Jumilla–Yecla region, SE Spain). Tectonophysics 265, 191–211.

Martín-Chivelet, J., Giménez, R., Luperto-Sinni, E., 1997. La discontinuidad del Campaniense basal en el Prebético ¿Inicio de la convergencia alpina en la Margen Bética? Geogaceta 22, 121–124.

Martín-Chivelet, J., Berástegui, X., Rosales, I., Vilas, L., Vera, J.A., Caus, E., Gräfe, K.-U., Segura, M., Puig, C., Mas, R., Robles, S., Floquet, M., Quesada, S., Ruiz-Ortiz, P.A., Fregenal-Martínez, M.A., Salas, R., García, A., Martín-Algarra, A., Arias, C., Meléndez, N., Chacón, B., Molina, J.M., Sanz, J.L., Castro, J.M., García-Hernández, M., Carenas, B., García-Hidalgo, J., Gil, J., Ortega, F., 2002. Cretaceous. In: Gibbons, W., Moreno, M.T. (Eds.), The Geology of Spain. Geological Society, London, pp. 255–292.

McCave, I.N., Hollister, C.D., 1985. Sedimentation under deep-sea current systems: Pre-HEBBLE ideas. In: Nowell, A.R.M., Hollister, C.D. (Eds.), Deep Ocean Sediment Transport—Preliminary Results of the High Energy Benthic Boundary Layer Experiment. Mar. Geol., vol. 66, pp. 13–24.

Nieto, L.M., 1997. La cuenca Subbética mesozoica en el sector oriental de las Cordilleras Béticas. Tesis doctoral. Universidad de Granada, Granada. 556 pp.

Nowell, A.R.M., Hollister, C.D., 1985. The objectives and rationale of HEBBLE. In: Nowell, A.R.M., Hollister, C.D. (Eds.), Deep Ocean Sediment Transport—Preliminary Results of the High Energy Benthic Boundary Layer Experiment. Mar. Geol., vol. 66, pp. 1–11.

Paquet, J., 1969. Étude géologique de l'ouest de la province de Murcie (Espagne). Mém. Soc. Géol. Fr., N.S. 48, 1–270 (Mém 111).

Philip, J., Floquet, M., 2000. Late Maastrichtian (69.5–65 Ma). In: Crasquin, S. (Ed.), Atlas Peri-Tethys, Palaeogeographic maps. Explanatory notes CCGM/CGMW, Paris, pp. 145–152.

Piper, D.J.W., Stow, D.A.V., 1991. Fine-grained turbidites. In: Einsele, G., Ricken, W., Seilacher, A. (Eds.), Cycles and Events in Stratigraphy. Springer, Berlin, pp. 360–376.

Poulsen, C.D., Seidov, E., Barron, W.P., 1998. The impact of Paleogeographic Evolution on the Surface Circulation and the Marine Environment within the mid-Cretaceous Tethys. Paleoceanography 13 (5), 546–559.

Reicherter, K.R., Pletsch, T.K., 2000. Evidence for a synchronous circum-Iberian subsidence event and its relation to the African–Iberian plate convergence in the Late Cretaceous. Terra Nova 12, 141–147.

Robaszynski, F., González Donoso, J.M., Linares, D., Amédro, F., Caron, M., Dupuis, C., Dhont, A.V., Gartner, S., 2000. Le crétacé supérieur de la région de Kalaat Senan, Tunisie centrale. Litho-Biostratigraphie intégrée: zones d'ammonites, de foraminifères planctoniques et de nannofossiles du Turonien supérieur au Maastrichtien. Bull. Cent. Rech. Elf Explor. Prod. 22 (2), 359–490.

Robin, E., Boclet, D., Bonté, P., Froguet, L., Jéhanno, C., Rocchia, R., 1991. The stratigraphic ditribution of Ni-rich spinels in Cretaceous-Tertiary boundary rocks at El Kef (Tunisia), Caravaca (Spain) and Hole 761 C (Leg 122). Earth Planet. Sci. Lett. 107, 715–721.

Savostin, L.A., Sibuet, J.C., Zonenshain, L.P., Le Pichon, X., Roulet, M.J., 1986. Kinematic evolution of the Tethys belt from the Atlantic Ocean to the Pamirs since the Triassic. Tectonophysics 228, 33–55.

Savrda, C.E., 1992. Trace Fossils and benthic oxygenation. In: Maples, C.G., West, R.R. (Eds.), Trace fossils. The Paleontological Society. Short Courses in Paleontology, vol. 5, pp. 172–196.

Shanmugam, G., 2000. 50 years of the turbidite paradigm (1950s–1990s): deep-water processes and facies models—a critical perspective. Mar. Pet. Geol. 17, 285–342.

Shanmugam, G., Spalding, T.D., Rofheart, D.H., 1993. Process sedimentology and reservoir quality of deep-marine bottom-current reworked sands (sandy contourites): An example from the Gulf of Mexico. AAPG Bull. 77 (7), 1241–1259.

Shanmugam, G., Spalding, T.D., Rofheart, D.H., 1995. Deep-marine bottom-current reworked sand (Pliocene and Pleistocene), Ewing Bank 826 Field, Gulf of Mexico. In: Winn Jr., R.D., Armentrout, J.M. (Eds.), Turbidites and associated deep-water facies, SEPM Core Workshop, vol. 20, pp. 25–54.

Simó, A., 1986. Carbonate platform depositional sequences, Upper Cretaceous, south–central Pyrenees (Spain). Tectonophysics 129, 205–231.

Stanley, D.J., 1993. Model for turbidite-to-contourite continuum and multiple processes transport in deep marine settings: examples in the rock record. Sediment. Geol. 82 (1/4), 241–256.

Stow, D.A.V., 1977. Late Quaternary Stratigraphy and Sedimentation on the Nova Scotian Continental Margin. PhD thesis, Dalhousie University, Canada.

Stow, D.A.V., 1982. Bottom currents and contourites in the North Atlantic. Bull. Inst. Geol. Bassin Aquitaine 31, 151–166.

Stow, D.A.V., Bowen, A.J., 1978. Origin of lamination in deep sea, fine-grained sediments. Nature 274, 324–328.

Stow, D.A.V., Faugères, J.C. (Eds.), 1993. Contourites and Bottom Currents. Sediment. Geol., vol. 82, pp. 1–310. Special issue.

Stow, D.A.V., Faugères, J.C. (Eds.), 1998. Contourites, Turbidites and Process Interaction. Sedimentary Geology, vol. 115, pp. 1–384. Special issue.

Stow, D.A.V., Lovell, J.P.B., 1979. Contourites: their recognition in modern and ancient sediments. Earth Sci. Rev. 14, 251–291.

Stow, D.A.V., Mayall, M. (Eds.), 2000a. Contourites, Turbidites and Process Interaction. Marine and Petroleum Geology, vol. 17, pp. 125–342. Special issue.

Stow, D.A.V., Mayall, M., 2000b. Deep-water sedimentary systems: New models for the 21st century. In: Stow, D.A.V., Mayall, M. (Eds.), Contourites, Turbidites and Process Interaction. Marine and Petroleum Geology, vol. 17, pp. 125–135.

Stow, D.A.V., Piper, D.J.W. (Eds.), 1984a. Fine-grained sediments: Deep-Water Processes and Facies. Gel. Soc. London, Spec. Publ., vol. 15, pp. 1–659.

Stow, D.A.V., Piper, D.J.W., 1984b. Deep-water fine-grained sediments: facies models. In: Stow, D.A.V., Piper, D.J.W. (Eds.), Fine-grained sediments: Deep-Water Processes and Facies. Gel. Soc. London, Spec. Publ., vol. 15, pp. 611–646.

Stow, D.A.V., Shanmugam, G., 1980. Sequence of structures in fine-grained turbidites: comparison of recent deep-sea and ancient flysch sediments. Sediment. Geol. 25, 23–42.

Stow, D.A.V., Faugères, J.C., Gonthier, E., 1986. Facies distribution and textural variation in Faro Drift contourites: velocity fluctuation and drift growth. Mar. Geol. 72, 71–100.

Stow, D.A.V., Reading, H.G., Collinson, J.C., 1996. Deep Seas. In: Reading, H.G. (Ed.), Sedimentary Environments: Processes, Facies and Stratigraphy, 3rd ed. Blackwell, Oxford, pp. 395–453.

Stow, D.A.V., Faugères, J.C., Viana, A., Gonthier, E., 1998. Fossil contourites: a critical review. In: Stow, D.A.V., Faugères, J.-C. (Eds.), Contourites, Turbidites and Process Interaction. Sediment. Geol., vol. 115, pp. 3–31.

Van Morkhoven, F.P.C.M., Berggren, W.A., Edwards, A.S., 1986. Cenozoic cosmo-politan deep-water benthic foraminifera. Bulletin des Centres de Recherches Exploration—Production Elf-Aquitaine Mem., vol. 11, pp. 1–423.

Van Veen, G.W., 1969. Geological investigations in the region west of Caravaca, southeastern Spain. Thesis. Univer. Amsterdam. 143 pp.

Vera, J.A., 1988. Evolución de los sistemas de depósito en el margen ibérico de la Cordillera Bética.. Rev. Soc. Geol. Esp. 1, 373–391.

Vera, J.A., Molina, J.M., 1999. La Formación Capas Rojas: caracterización y génesis. Estud. Geol. 55, 45–66.

Vera, J.A., García-Hernández, M., López-Garrido, A.C., Comas, M.C., Ruiz-Ortiz, P.A., Martín-Algarra, A., 1982. El Cretácico de la Cordillera Bética. In: Garcia, A. (Ed.), El Cretácico de España. Universidad Complutense, Madrid, pp. 515–632.

Viana, A.R., Faugères, J.-C., Stow, D.A.V., 1998. Bottom current controlled sand deposits—a review from modern shallow and deep water environments. In: Stow, D.A.V., Faugères, J.-C. (Eds.), Contourites, Turbidites and Process Interaction. Sediment. Geol., vol. 115, pp. 53–80.

Wezel, F.C., 1969. Prossimalita, distalita e analisi dei bacini dei flysch: un punto di vista attualistico. Mem. Soc. Natl. Napoli, Suppl. Boll. 78, 1–12.

Wüst, G., 1936. Schichtung und Zirkulation des Atlantischen Ozeans. Das Bodenwasser und die Stratosphäre. Wiss. Ergeb. ber - Dtsch. Atlant. 6, 1–288 (Meteor 1925–27).

Wüst, G., 1958. Die Stromgeschwindigkeiten und Strommengen in der Atlantischen Tiefsee. Geol. Rundsch. 47, 187–195.

Yokokawa, M., Kishi, M., Masuda, F., Yamanaka, M., 1995. Climbing ripples recording the change of tidal current condition in the Middle Pleistocene Shimosa Group, Japan. In: Fleming, B.W., Bartholomä, A. (Eds.), Tidal Signatures in Modern and Ancient Sediments. International Assoc. Sediment. Spec. Publ., vol. 24. Blackwell Science, Oxford, pp. 301–312.