Salt diapirs are probably one of the most complex geological structures affecting sedimentary rocks. They are formed by salt bodies with a very variable shape (including such features as fingers, bubbles and overhangs) that deform the overburden, producing a complex set of folds and faults (Jackson & Talbot 1991; Jackson 1995; Stewart 2006; Hudec & Jackson 2007). Consequently, the availability of geophysical data is crucial for a correct characterization of these structures, the geometry of which changes drastically both laterally and with depth. In such a complex geological setting, acquiring geophysical data of suitable quality to allow a reasonable interpretation and validate the structural concepts is fundamental. The available geophysical techniques play different roles from initial exploration to drilling (Henke et al. 2005). Potential field methods (gravity, magnetic) are more suitable during the initial steps, whereas seismic methods (seismic reflection, wide-angle seismic refraction) are the most widespread technique used during the final prospect and drilling. Gravity methods allow us to define the geometry of the diapir body from the density differences between the diapirc rocks (salt) and the overburden (detrital and carbonate rocks) (Nettleton 1971; Nagihara & Hall 2001; Jacques et al. 2003), but not the internal structure of the overburden. Furthermore, this technique, given its inherent ambiguity, allows different interpretations. Magnetic methods are less used on diapir studies although high-sensitivity magnetic surveys allow researchers to map salt domes and salt ridges as magnetic lows (Baumgartner & Van Andel 1971; Al-Zoubi & ten Brink 2001; Pearson 2006) or exceptionally, if diagenetic magnetite is present within the anhydrite and halite, as positive magnetic anomalies (Smith & Whitehead 1989). However, this technique is also ambiguous in terms of the precise shape and size of the salt structures. The seismic reflection technique usually does not have these two limitations. It provides a detailed image of the roof of the diapir body as well as the internal structure of the overburden (Rowan & Vendeville 2006; Stewart 2006). Using wide-angle reflection and refraction data, as well as common depth point (CDP) processing, allows us to image the geometry of the entire diapir body and that of the underlying rocks (Malinowski et al. 2007). However, conventional seismic reflection cannot accurately characterize vertical limits, such as diapir walls and surrounding drag folds with near-vertical limbs. Moreover, for the present, it is an expensive technique and is difficult to use onshore because of both logistical (i.e. regions with complex topography) and administrative difficulties.

In this setting, the magnetotelluric (MT) method may fill the gap between the potential field methods and the seismic techniques. The MT method allows us to obtain an image of the electrical resistivity (or its inverse, the electrical conductivity) distribution of the subsurface. The wide range of electrical resistivity in rocks is related to various factors such as porosity, water content, salinity and mineral composition. As diapirs are usually formed by salt (which has high resistivity values when dry) and the overburden by porous sedimentary rocks (which are less resistive), MT is a suitable method for prospecting salt diapirc structures and characterizing their geometry, as well as that of its substratum and overburden. Also, this method can allow us to determine whether the porous sedimentary rocks of the overburden are filled by conductive fluids (i.e. saline water) or resistive fluids (freshwater, gas or oil). Thus, some studies have modelled the shape of salt bodies (Hoversten et al. 2000; Key et al. 2006), and even that of sub-salt structures (Newman et al. 2002) using the MT method; however, most of these are offshore studies that give only a broad image of the diapiric structures because of the spacing and number of stations recorded.

The main aim of this study is to determine the geometry of a salt diapir using the MT method with a high density of measurement sites. Also, we want to investigate the extent to which the
MT method can be used to record fluid and overburden lithology changes. To achieve this twofold purpose, we have recorded 34 MT measuring stations across the 13 km long Bicorp–Quesa Diapir profile, whose surface structure has been well defined from field mapping and structural measurements (mainly bedding and fault attitudes). This diapir is located in the most external part of the Prebetic Zone (SE Spain). It is an elongate diapir in which both internal and overburden structures are parallel to the diapir walls, extending without significant orientation changes for 11–12 km (Rios et al. 1980; De Ruig 1992; Roca et al. 1996). Therefore, it is a structure with smooth longitudinal changes, which permits its MT characterization by means of a transverse profile.

Geological setting

The Prebetic Zone is the most external part of the foreland fold and thrust belt of the Betic Cordillera. It consists of a para-autochthonous Mesozoic to middle Miocene cover detached from the Variscan basement on Triassic evaporite and mudstone layers (Vera 1983). The Mesozoic to Cenozoic (middle Miocene) cover is deformed by a series of folds, faults and diapirs composed of Triassic evaporite–mudstone that are mainly oriented ENE–WSW but also, locally, NNW–SSE (De Ruig 1992; Roca et al. 2006; Fig. 1). These structures reveal a complex deformation history with an early to middle Miocene contractional deformation and, locally, affected by a younger late Miocene extension (Ott d'Estevou et al. 1988; De Ruig 1995; Roca et al. 1996). The contractional deformation is predominant and decreases significantly to the NNW. On the basis of this decrease in the degree of contractional deformation, three structural domains can be distinguished in the eastern Prebetics (Aznárez et al. 1979; García-Hernández et al. 1980; De Ruig 1992), as follows.

1. The strongly deformed Internal Prebetic Zone, located to the south, shows a complex structure including NNW-directed thrusts and ENE–WSE- to east–west-trending strike-slip faults, as well as folds, diapirs and normal faults with variable orientations (Fig. 1).

2. The External Prebetic Zone is characterized by ENE–WSW-trending box anticlines separated by broad synclines and cut by oblique to transverse normal faults (Figs 1 and 2). The anticlines are cored by Middle–Late Triassic evaporite and mudstone, and their northern limbs are often overturned and thrust over the adjacent synclines (García-Rodrigo 1960; De Ruig 1992). Most of these anticlines also include squeezed diapirs on their crests. The development of these External Prebetic contractional structures is related to the moderate positive inversion of a pre-existing horst-and-graben system pierced by salt diapirs (De Ruig 1992; Martinez del Olmo 1999; Roca et al. 2006).

3. The Valencian Domain is characterized by a large, 40 km wide thrust sheet displaced less than 2 km to the north (Roca et al. 2006), located between the Iberian Chain and the External...
Prebetic Zone. This thrust sheet consists of a central subhorizontal platform of 1000–1500 m thick Jurassic–Cretaceous carbonates (mainly dolostone and limestone) bounded by NNW–SSE- and ENE–WSW-trending folds and thrusts at its eastern and northern boundaries, respectively (Fig. 2). The platform as well as these folds and thrusts are cut by two nearly orthogonal sets of normal faults that also strike ENE–WSW and NNW–SSE. These contemporaneous normal faults bound a complex system of narrow grabens that were pierced during the Miocene by elongate diapirs of Triassic evaporite and mudstone (Moissenet 1985, 1989; Figs 1 and 2). Usually, the grabens are filled by Miocene continental to lacustrine sediments that have growth strata geometries close to the diapir walls (Santisteban et al. 1989; Roca et al. 1996; Anadón et al. 1998). One of the salt diapirs along the axis of these grabens is the Bicorb–Quesa Diapir, which is the subject of this study.

**Bicorb–Quesa Diapir**

This diapir is located along the axis of the Bicorb–Quesa graben system, which crosses the central part of the subhorizontal platform of the Valencian Domain in an ENE–WSW direction. It is 12 km long and 1 km wide (Fig. 3) and is mainly composed of Upper Triassic continental deposits (Keuper facies). These deposits include: a Lower Evaporitic Unit, formed by a salt layer overlain by grey gypsiferous marls, carbonates and sandstones; a red Mid-Detrital Unit, composed of sandstones and mudstones; and an Upper Evaporitic Unit, formed by salt, gypsum and red gypsiferous mudstones (Orti 1974; Bartrina et al. 1990; De Torres & Sánchez 1990). Besides these evaporite and mudstone layers, in the western part, there are also Middle Triassic carbonates (Muschelkalk facies) that form a major slab on the top of the outcropping diapir (Fig. 3).

The Bicorb–Quesa Diapir is flanked by two parallel half-grabens filled by syn-diapiric Miocene strata (Roca et al. 1996; Anadón et al. 1998). The northern one, the Bicorb Half-graben, has a thicker basin fill (up to 650 m; Santisteban et al., 1989; Anadón et al. 1998) and is formed by a block tilted 20–60° northward bounded on the north by two SSE-dipping normal faults. These faults accumulate a vertical throw of 1000 m and separate the Bicorb Basin from a subhorizontal Jurassic–Cretaceous platform (here called the North Platform), which appears

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**Fig. 3.** Geological map of Bicorb–Quesa graben system with the piercing Bicorb–Quesa Diapir. B–B’, MT modelled cross-section; white dashed box shows location of Figure 4.
slightly tilted to the north close to both faults (Fig. 3). The other side of the diapir is flanked by the Quesa Half-graben. The structure of this half-graben is more complex, as it is compartmentalized into several south-dipping blocks bounded by normal faults. Also, it has a thinner basin fill (less than 450 m) and, to the south, it is bounded by a NNW-dipping normal fault with a throw that does not exceed 600 m. South of this fault, the Jurassic–Cretaceous sequences are again subhorizontal, forming a large platform (here called the South Platform), which extends to the External Prebetic. In this South Platform, the outcropping Upper Cretaceous units are thinner and are uplifted in relation to the equivalent units in the platform located north of the Bicorb Half-graben. According to Roca et al. (1996), these thickness and altitude variations suggest the pre-existence of a north-dipping Mesozoic normal fault along the Bicorb–Quesa Half-graben system.

MT method

The MT method is a natural-source electromagnetic (EM) technique that allows us to characterize the electrical conductivity distribution of the subsurface by measuring simultaneously the time variation of the electric and magnetic fields on the Earth’s surface. The penetration is a function of the electrical conductivity of the Earth (\(\sigma\)) and the frequency (\(f\)) of investigation. A reasonable measure of inductive scale length is given by the skin depth (\(d \approx 500(\sigma f)^{1/2}\) m if \(f\) is in Hz and \(\sigma\) is in S m\(^{-1}\)), which is the depth by which the incident EM fields have attenuated to \(1/e\) of the surface value in a uniform space. This technique was originally presented by Tikhonov (1950) and Cagniard (1953), and has become a useful tool for characterizing the geoelectrical structure at a wide range of depths from a few metres to the mantle, depending on the recording frequency. The fundamentals of the method have been extensively presented by Vozoff (1991) and Simpson & Bahr (2005).

Magnetotelluric data and 2D inversion

A total of 34 measuring stations (sites) have been recorded in the study area using broadband ADU-06 MT stations from Metronix. The measured frequency ranged between \(10^{-3}\) s and \(10^2\) s. Taking in account the rather constant ENE–WSW trend of the geological structures and assuming that the strike of the geoelectrical bodies will be the same, the sites were located along a NNW–SSE profile (Fig. 4). It was attempted to keep the distance between the sites and the profile trace at a minimum, except on the subhorizontal platforms, where the choice of best emplacement conditions (i.e. vast flat topography, easy access) prevailed. The location and spacing of the sites was chosen according to the structure in such a way that five sites were placed on the outcropping diapir and at least two on each recognized overburden block that is bounded by a major fault or by the diapir. Consequently, the average site density is higher in the centre of the model (<200 m) than on the less deformed platforms, where it is of the order of 500–1000 m.

The MT dimensionality of the acquired data has been defined using the WALDIM software (Martí et al. 2004). This software calculates the invariant rotational values of the impedance tensor established by Weaver et al. (2000) as well as their errors. The analysis of the dimensionality shows a regional 2D response within the range of 0.001–100 s and with some local distortion in some cases (3D or 2D). For 2D regional geoelectric structure, the modelling processing could be decoupling in two modes: one mode is descriptive of the responses for electric currents flowing
along the structure (TE mode), and the other for currents flowing across the structure (TM mode).

To determine the direction of the geological structures (strike) and the TE and TM modes we applied the McNeice–Jones (McNeice & Jones 2001) multi-site, multi-frequency MT tensor decomposition based on the study by Groom & Bailey (1989). The multi-frequency geoelectric strike directions are shown in Figure 4 for the period band 0.001–100 s. These broad period bands yield the preferred strike of the bulk of the upper crustal bodies, and are dominated by the response of regional large-scale structures. The lengths of the strike arrows are scaled by the normalized root mean square (r.m.s.) misfit error between the observed and the estimated impedance. The data can be considered 2D with a geoelectric strike direction close to N63°E, which is parallel to the main geological features.

Joint 2D inversions of the TE and TM apparent resistivity and phase data were undertaken using the algorithm of Rodi & Mackie (2001). The algorithm searches for the model that trades off the lowest possible misfit against the lowest possible smoothness, measured by the sum of the lateral and vertical gradients in resistivity. The error floor adopted for the impedance tensor data was 5%; the model responses fit the apparent resistivity data to within 10%, and the phases to within 2.8°, on average. Comparisons between the data and model responses are shown in Figure 5 for the apparent resistivities and phases of both modes. The final model derived is shown in Figure 6.

Results

The high-quality outcrops in the study area allow us to constrain the structure and thickness of the Cretaceous and Cenozoic differentiated units at the shallower part of the MT profile. This is well depicted in the detailed cross-section shown in Figure 6a. The main features visible in the cross-section are as follows: (1) north of the diapir there is southward thinning of the Cenomanian–Turonian units whereas south of it the thickness of these
units is constant; (2) the Quesa Half-graben is compartmentalized, with several synthetic faults (one of them reverse); (3) the main fault of the Bicorb Half-graben is fossilized by the upper Miocene units; (4) the lower Miocene units erode the Senonian sequences. Therefore the structure at the surface is well defined by the outcropping units. However, it is not clear if this structure changes with depth and the thickness of the non-outcropping units is not known.

**Origin of the observed resistivity changes**

The inversion of the acquired data denotes significant resistivity contrasts in the bedrock of the study area. Over a deep resistive basement, the model shows a high-resistivity body overlaid by a shallow very high-conductivity zone in the middle of the profile (diapir area), and, on both sides of this central resistive body, an alternation of resistive and conductive layers (Fig. 6b). These layers are near-horizontal and continuous beneath the platforms, whereas beneath the Bicorb and Quesa half-grabens they are discontinuous (and consequently less defined) and tilted towards the SSE (Quesa Half-Graben) or NNW (Bicorb Half-graben).

**Overburden and basement interpretation.** Overprinting the MT model with the geological cross-section obtained from field data (Fig. 6b) highlights the good correlation between resistivity and lithology in the overburden (Fig. 7). Thus, outcropping dolostones and marly dolostones (Cenomanian to Turonian in age) show a high-resistivity response in the model, whereas mudstones, sandstones and calcarenites (Early Cretaceous and Miocene in age) appear as conductive bodies or layers (Figs 6b and 7). The limestone layers (Senonian and Miocene) are resistive except in their lowermost parts and close the diapir where they may be conductive; this conductivity could be related to fluid circulation over the contact of these limestones with the underlying less permeable rocks (mudstone or dolostone).

Taking into account this resistivity–lithology correlation and the known regional lithostratigraphy, the highly resistive level located beneath the outcropping Lower Cretaceous conductor has
been linked to the Jurassic and the underlying conductor to the Upper Triassic units (Keuper facies) (Fig. 7). Indeed, regional studies (Ortí 1974; Rios et al. 1980; Lanaja 1987; Bartrina et al. 1990; Sopena et al. 1990) show a 400–600 m thick Jurassic sequence of dolostones and limestones, which could correlate very well with the 300–600 m thick high-resistivity structure; and the Triassic units, mainly mudstones, evaporites, sandstones and limestones, can be correlated with the deep low-resistivity structure.

Beneath this alternation of resistant and conductive layers linked to lithology changes in the Mesozoic cover and overlying Miocene sediments, the high-resistivity basement has been correlated with the Palaeozoic Variscan basement (Figs 6b and 7). This basement is composed of metamorphic sandstones, shale and quartzites (Bartrina et al. 1990; Martínez-Poyatos et al. 2004), and shows a high-resistivity response in the Iberian Massif (Muñoz et al. 2008), Pyrenees (Ledo et al. 1998), Betic Cordillera (Martí et al. 2009) and Catalan Coastal Ranges (M. A. Marin, pers. comm.).

In summary, the obtained MT model shows that the lithology is the main factor controlling the resistivity changes observed for the diapir. Thus, the highest resistivity values show a good correlation with the basement metamorphic rocks and dolostones, the medium resistivity values with the limestones, and the lowest values with mudstones, sandstones and evaporites (gypsum and anhydrite) (Fig. 7). In relation to this correlation, higher resistivity values (A in Fig. 6b) are observed in the Jurassic materials under the North Platform, as in the Turonian to Senonian carbonate levels located under the Bicorb Half-graben. These resistivity values suggest that there they are mainly related to dolomites that are more resistive than limestones (Telford et al. 1976).

Finally, it should be noted that, for the diapir, the proposed resistivity–lithology correlation is not always clear. There are two zones in which high and low resistivity values occur whose correlation with lithological changes is not clear (structures C and D in Fig. 6b). Both zones are located in the Quesa Half-graben, where the overburden is compartmentalized by a complex system of normal faults with displacements that change quickly along their trace (Fig. 3). Also, they are related to MT sites that are located far from the profile trace (sites a and b in Figs 4 and 6b, 500 m and 200 m from the profile, respectively) and that show a less marked geoelectric strike direction (Fig. 4, sites a and b).

### Bicorb–Quesa structure

The lithostratigraphic image offered by the MT model allows us to improve substantially the knowledge of the subsurface structure of the Bicorb–Quesa Diapir and surrounding overburden areas, and provides information that cannot be derived from geological surface data. Thus, the MT model (Fig. 6) shows the following features.

1. The top of the Variscan basement is located at a different depth north and south of the Bicorb–Quesa Diapir. Whereas beneath the North Platform and Bicorb Half-graben it is located c. 2500 m below sea level, beneath the South Platform and the Quesa Half-graben it appears at a shallow depth (c. 1500 m).

2. Outcropping faults affecting the overburden correlates at depth with sudden changes of the resistive pattern of the MT model. Specifically, they coincide with the lateral disappearance of high- and low-resistivity structures. This is well depicted beneath the Quesa Half-graben, where the low-resistivity structures linked to the Lower Cretaceous units disappear at the deep prolongation of both the master fault bounding the half-graben to the south and the faults located in the axis of this half-graben.

3. The Jurassic–Cretaceous overburden is tilted close the
Bicorb–Quesa Diapir, and is much more tilted on the north flank of the diapir than on the south one.

(4) The Bicorb–Quesa Diapir is noticeably asymmetrical, with the south wall dipping c. 50° to the south whereas the north wall is nearly vertical.

Apart from providing these intrinsic improvements in our knowledge of the deep structure of the region, these features indicate that the Bicorb–Quesa Diapir is located above a normal basement fault (Bicorb–Quesa Fault) that lowered the NNW block and was active during the Mesozoic. This fault was already suggested by Roca et al. (1996) from the different thickness and elevation shown by the Cenomanian and Turonian on the North and South platforms; however, the MT data allow us to define its location, displacement and age of motion (Triassic? to Turonian).

The presence of normal faults active during Triassic and Late Jurassic–Cretaceous times has been described in the regions surrounding the Valencian Domain (Sopeña et al. 1990; De Ruig 1992; Peper & Cloetingh 1992); however, most of them were inferred from stratigraphic thickness differences and therefore they are faults of unknown geometry and imprecise location.

The tilted shape of the Bicorb–Quesa Diapir could denote the shortening of a previous diapir (e.g. Letouzey et al. 1995; Canérot et al. 2005; Roca et al. 2006; Hudec & Jackson 2006) or a surrounding asymmetric deformation or asymmetric deposition generated during the withdrawal or salt rise controlled by overburden structures such as asymmetric grabens (Schultz-Ela 2003). In the present case, the tilting of the Bicorb–Quesa Diapir seems to be related to a combination of both causes: (1) the shortening of a early Miocene diapir during the middle Miocene Betic compression (Roca et al. 1996; Anadón et al. 1998); (2) the deposition and deformation asymmetry generated by the motion of the Bicorb–Quesa basement fault, which produced much more deformation and syn-diapir sedimentation on the north than on the south side of the diapir.

Discussion

From the results obtained, it is evident that the MT method is a suitable and useful technique to characterize not only the deep geometry of diapirs but also the geometry of the adjoining overburden. Also, it is a powerful tool to decipher the motion and migration of the salt rocks inside the diapiric rocks. If we take into account that dry salt is highly resistive, the MT method allows us to elucidate where it accumulates and where it disappears. For example, in the Bicorb–Quesa Diapir, we observe that the diapiric rocks (Keuper facies) are highly resistive only in the core of the diapir whereas the rest of the diapir is conductive. This indicates that all the salt has migrated from the lowermost part of the platforms and half-grabens to the diapir and, consequently, that, at present, the Bicorb–Quesa Diapir cannot grow because of lack of salt supply.

On the other hand, the MT study also provides information about the fluid nature and circulation in and around the diapirs. In the present case, it shows that saline waters are restricted to the top of the diapir and that they do not encroach into the surrounding overburden. Also, it indicates fluid circulation of slightly conductive waters at the bottom of the limestone formations.

Conclusions

The MT method has been proved to be a powerful and useful tool to characterize not only the geometry (structure) of the diapirs, underlying rocks and adjoining overburden but also the nature of the fluid circulating in these rocks. Also, it has been observed that it is a method that can be employed to decipher salt accumulations in the diapiric rocks and, therefore, to understand salt flow kinematics. Specifically, in the Bicorb–Quesa region, the MT technique has allowed us to determine various features as follows.

(1) The geometry of the diapir and overburden has been characterized from the established correlation between lithology and resistivity. In particular, comparison between the MT model and field data has allowed us to correlate: (a) the conductors with stratigraphic levels mainly composed of mudstones and sandstones; (b) the overburden resistive layers with dolostones and limestones; (c) the high-resistivity diapir core with dry salt.

(2) The top of the Variscan basement is a resistive basement that is affected by a north-dipping normal fault. This fault is located beneath the Bicorb–Quesa Diapir and has a vertical throw of 1000 m. It was active during the Mesozoic, as shown by the different thicknesses of sequences to either side of the fault, and by the resistive and conductive levels correlated with the Mesozoic major lithostratigraphic units.

(3) The shape of the diapir is asymmetrical, with a north wall that is vertical and a south wall dipping southwards.

(4) All the Keuper salt has been withdrawn to the diapir in such a way that it constitutes the predominant lithology in the diapir core whereas below the platforms it is absent and the Keuper facies appears to be formed by conductive mudstones and sandstones.

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