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ORIGINAL PAPER

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Tertiary tectonics of the Dannemarie Basin, upper Rhine graben, and regional implications

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Abstract Recently released seismic reflection data, together with previous seismic and well data, are used to describe the development of the Dannemarie basin, in the SW end of the Upper Rhine Graben. The Dannemarie Basin was formed during the main rifting phase of the Upper Rhine Graben as an asymmetrical graben trending NE–SW. Post-rift tectonism shifted the depocenter southward and changed the overall shape of the basin. Miocene Jura compression did not result in the formation of folds, as in the adjacent Mulhouse Horst. Strike slip faulting was dominant in the post-rift period and new faults were created, most notably the north trending and transpressional Belfort Fault. The boundary of the Dannemarie Basin with the Vosges Mountains is part of a restraining bend, which may account for the uplift of the southernmost part of the Vosges Mountains.

Keywords Rhine Graben · Dannemarie Basin · Vosges Mountains · Seismic reflection

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Introduction

The Dannemarie Basin (DB) occupies the southwestern corner of the Upper Rhine Graben (URG), between the Vosges Mountains and the Tabular Jura (Fig. 1). To the east it is bounded by the Mulhouse High (Mulhouse Horst, MH) along the Illfurth Fault. On the west it borders the Saône Transfer Zone, which separates the URG from the continuation of the central European rift system in the Bresse Graben (Ziegler 1992). This location, in the interface between several tectonic units, makes the DB potentially important for the understanding of the past and present regional tectonics. Open questions include the structure of the area before the onset of Miocene folding in the Swiss and French Jura, and the role of Jura compression in shaping the present structure of the URG. In addition, although the Saône Transfer Zone was discussed in several works (e.g. Contini and Théobald 1974; Illies 1977; Bergerat and Chorowicz 1981; Lacombe et al. 1993), its relation through time with the URG is not fully resolved. Finally, a puzzling problem is the rise of the southern Vosges Mountains that abuts the DB on the north, to their present height. It was suggested to be a rift shoulder uplift (Villemin et al. 1986; Villemin and Bergerat 1987), but is commonly thought to be a Pliocene or younger phenomenon (e.g. Illies 1972, 1975; Ziegler 1994; Duringer 2001) and is, thus, not likely to be explained only by conventional rift shoulder uplift.

The DB was the focus of hydrocarbon exploration, which included seismic surveys and a number of exploration wells (Fig. 2) that revealed its overall structure (Doebl and Olbrecht 1974; Sittler 1974). However, the amount and quality of these older seismic data were, for the most part, insufficient for a detailed study of the basin. In recent years, two high quality seismic data sets that cover a large part of the DB with closely spaced lines, acquired in 1985 and in 1987 became available. This data set was never discussed in the literature, but was previously studied by Dupont (2000) and by

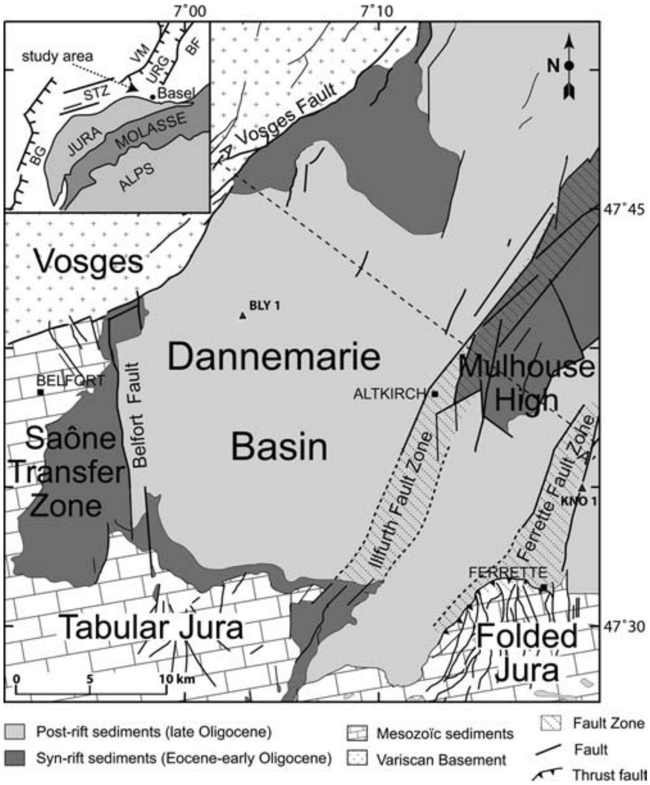


Fig. 1 Simplified geologic map of the Dannemarie basin and adjacent areas. *BLY1* Bellemagny; *KNO1* Knoeringue well

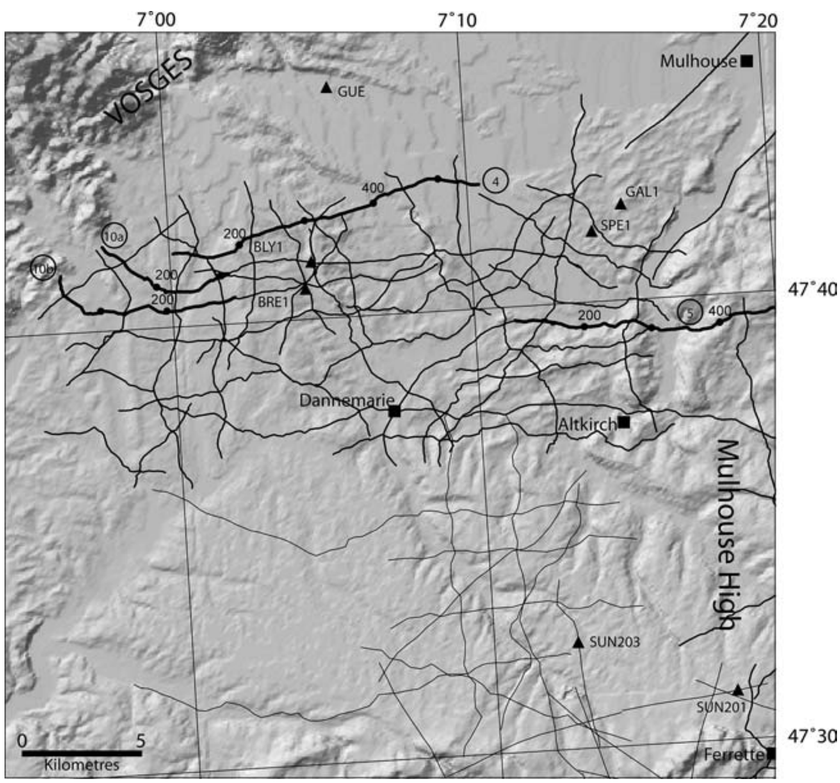
Bourgeois et al. (2001) as part of the GeoFrance3D program. Their study concentrated on achieving an accurate shape of the present basin by 3D depth migration. In this work, we use this set of seismic lines, together with wells and older lines from the northern and southern parts of the DB-areas that were not properly sampled by the more recent seismic surveys. We present depth and isopach maps and use them to discuss the main stages in the development of the DB and its role in the regional Tertiary geology. In particular, we examine the data in terms of the unresolved issues in the regional geology.

Geologic setting

The URG is part of the European Cenozoic rift system, which extends for over 1,100 km from the Mediterranean to the North Sea (e.g. Ziegler 1992). It was the target of numerous geological and geophysical investigations for over a century and the results were summarized in several reviews (Illies 1974; Pflug 1982; Hűttner 1991) and more recently by Sissingh (1998, 2003), Lutz and Cleintuar (1999) and Schumacher (2002).

Stratigraphic information is available from the several wells in the area (Fig. 2) and is summarized in Fig. 3. The deepest well in the area is Knoeringue 1 (Fig. 1, KNO 1) that reached basement at a depth of 2,150 m. The well penetrated only some 275 m of Tertiary sediments, compared to about 1,500 m, Tertiary

Fig. 2 Location map of seismic lines and wells used in this work, superimposed on a digital terrain map. Seismic lines that are shown in figures are *bold*, with figure number. *Intermediate thickness* denotes 1985 and 1987 vintages and *thin lines* denote early seismic lines. Wells are: *BLY1* Bellemagny 1D; *BRE1* Brechaumont 1; *GAL1* Galfingue 1; *GUE* Guewenheim; *SPE1* Spechbach; *SUN201* Sundgau 201; *SUN 203* Sundgau 203



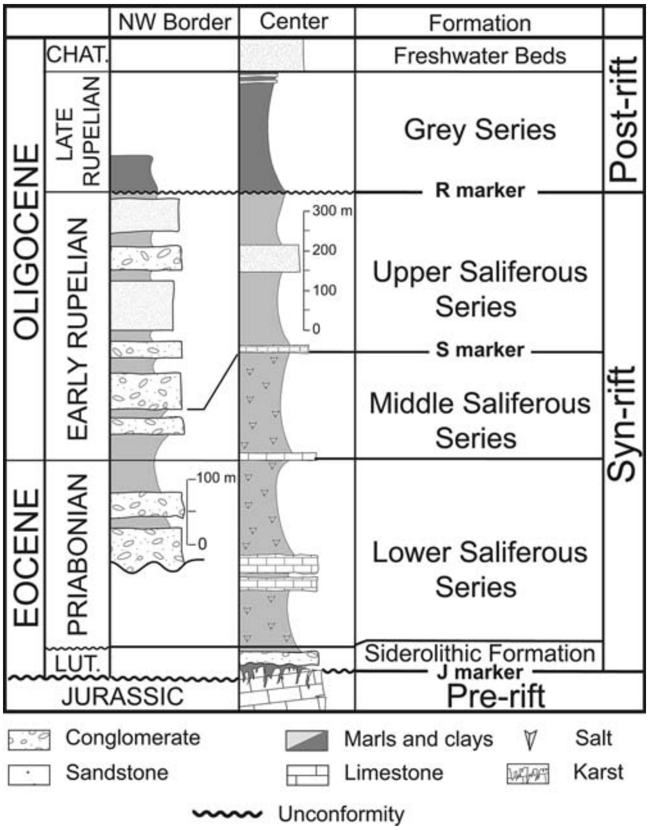


Fig. 3 Simplified stratigraphic column of the Tertiary in the DB at its center (*BLY 1*), and at the northern part (Guewenheim well, Jung and Schneegans 1930). Also shown are the stratigraphic locations of the markers that are used in this work. For the locations of the wells see Fig. 2

sediments in the DB. Pre-rift sediments in the URG are Mesozoic platform sediments that include an unusually thick Jurassic section of some 700 m, which is exposed in the Tabular Jura. The Jurassic is underlain in the KNO 1 well by Triassic sequences, including the Muschelkalk formation with thin veneers of salt. Based on the KNO 1 well, Permian rocks are likely to overlie the basement, but the wells in the DB were not deep enough to penetrate the Permian.

The top of the Jurassic in the southern URG is an unconformity overlain by marls and conglomerates of the Lutetian Siderolithic Formation in the center of the DB (Sissingh 1998), or by Priabonian conglomerates at the NW part of the basin (Sissingh 2003), considered as the earliest graben deposits (Fig. 3). The main phase of rifting and Tertiary deposition in the southern URG is represented by early Rupelian (Sannoisian) sandstones, limestones and marls (Sissingh 1998; Schumacher 2002). These rocks, mostly accumulated in a lacustrine environment (Duringer 1988), are usually divided into three main depositional sequences: lower, middle and upper Saliferous Zones.

Following the end of the main rifting phase of the southern URG in middle Rupelian (e.g. Schumacher 2002; Sissingh 2003), it subsided thermally, except for a period of pronounced uplift in middle Miocene (Roll 1979). This uplift was accompanied by an isolated burst of volcanism in the Kaiserstuhl area, about 50 km NE of the DB (Wimmenauer 1966). Post-rift sediments include middle and late Rupelian and, at places, Chattian. However, Chattian sediments are mostly absent in the southern URG, together with some of the underlying Rupelian rocks, due to erosion induced by subsequent Miocene uplift (Sissingh 1998; Schumacher 2002). This uplift and associated erosion is considered the result of Alpine compression (Sinclair et al. 1991; Laubscher 1992), mantle uplift (Illies 1977; Villemin et al. 1986) and at places transpressional reactivation of the Saône/Burgundy Transform Zone (Schumacher 2002).

Seismic data

The two industry seismic surveys that are mainly discussed in this work were collected using identical parameters. Geophone spread of 65–70 m to 1,350 m enabled the acquisition of data with good velocity control, from near the surface to the base of the approximately 1,500 m deep Tertiary graben. In many of the lines, high quality record is available also from the Mesozoic sequences. Station interval of 30 m and a vibrator frequency band of 12–96 Hz also assured a relatively high lateral and vertical resolution. Thus, in general, local conditions and optimal field parameters resulted in a set of high quality seismic sections. Standard commercial land processing was applied to the data at the time it was collected.

Two seismic markers were mainly used in this work, the J and the R markers (Fig. 4). The J marker is associated with the top Jurassic/base Tertiary erosional unconformity at the onset of rifting, and is identified by the unconformity that is associated with it (Fig. 5). The R marker is associated with the middle Rupelian transgressional unconformity at the end of the main rifting phase in the southern URG (Sissingh 2003). R is a regional and easily recognizable seismic marker in the southern URG (see Lutz and Cleintuar 1999). In addition, depths and velocities are available from the BLY 1 well that was drilled on one of the seismic lines (Fig. 2), and are consistent with these determinations. Also shown in the sections is the S unconformity providing separation between the lower and middle saliferous sequences that include the evaporites, and the upper saliferous sequence without evaporites. Several pre-rift Mesozoic seismic stratigraphic markers are shown in some of the sections, but they were also not mapped. These additional seismic markers are shown for the sake of completeness. Their identification is based on the depth and velocity information from the BLY-1 well. On the MH, additional depth information is available from the KNO 1 well that is used to constrain the two way time of the markers, using the

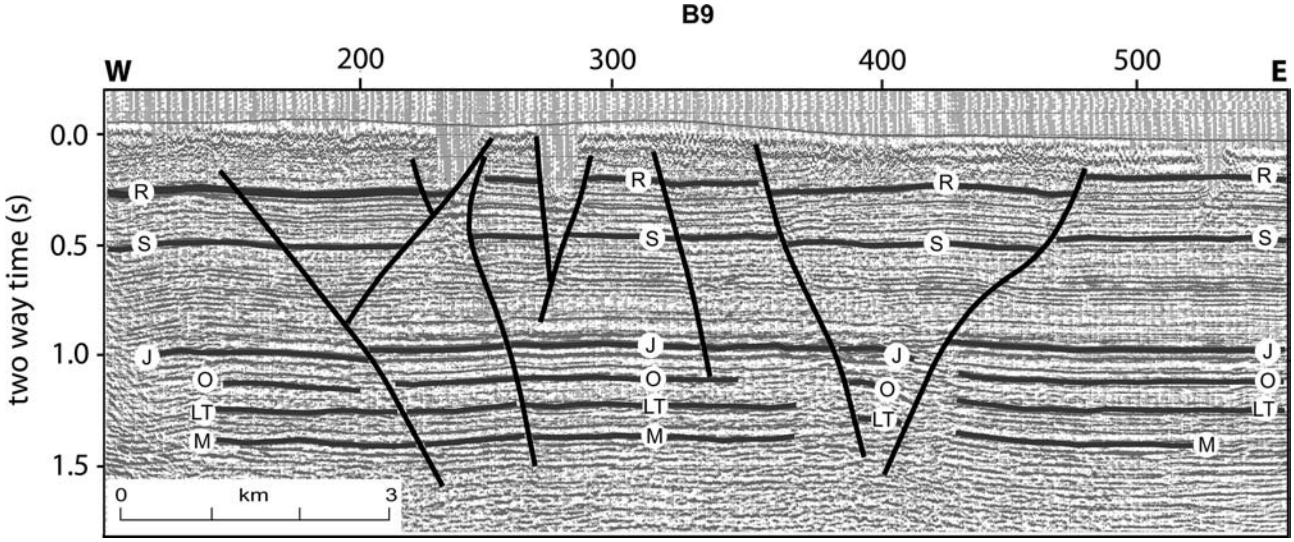


Fig. 4 Seismic section B9. For location see Fig. 2. *R* Mid Rupelian, marking the end of the main rifting phase. *S* Boundary between upper saliferous sequence to the middle saliferous sequence. *J* Top Jurassic/base Tertiary unconformity and onset of rifting. *O* Top Grande Oolithe formation (Jurassic). *LT* Near top Liassic, or top Triassic. *M* Top Mushelkalk Formation (Triassic)

BLY 1 velocities. However, the J and R markers are also exposed on the surface, in the southern and northern parts of the area, respectively. In both the DB and the MH, the seismic markers are also easily identified by the unconformities associated with them or, for the Mesozoic markers, by the laterally constant appearance of the reflector sequence throughout both the DB and MH. Depths were derived by simple 1D depth conversion, using average velocities of 2,200 m/s for the post-rift sediments and 3,500 m/s for the rift period sediments, based on the data from the BLY 1 well.

Older seismic data are also available from the area (Fig. 2). These are also usually 24 fold sections, but are of a much more variable quality in terms of reflector continuity and resolution. The older data were used mainly to extend the maps southward, beyond the extent of the more recent surveys. Although some of the older lines in this additional area to the south show clearly the seismic markers, we treat this part as being approximate.

Pre-rift-record

Many of the seismic sections include information on pre-rift Mesozoic and, at times, Paleozoic strata. The seismic data show that throughout the DB, the Mesozoic section is quite uniform. Depth to basement in the DB can be estimated from the observed depth to the top Jurassic, by adding some 1,900 m, the observed thickness of the Mesozoic and Paleozoic section in the adjacent KNO 1 well. Thus, based on the KNO1 data and Tertiary thickness (Doebl and Olbrecht 1974), the top of the crystalline basement in the center of the DB is likely to be over 3,500 m below the surface. However, as the thickness of the Paleozoic that is recorded in the KNO 1 well may vary laterally, this figure may change somewhat.

Rift period structures

The main features of the DB were formed during the Eocene–Oligocene rifting phase of the URG. In order

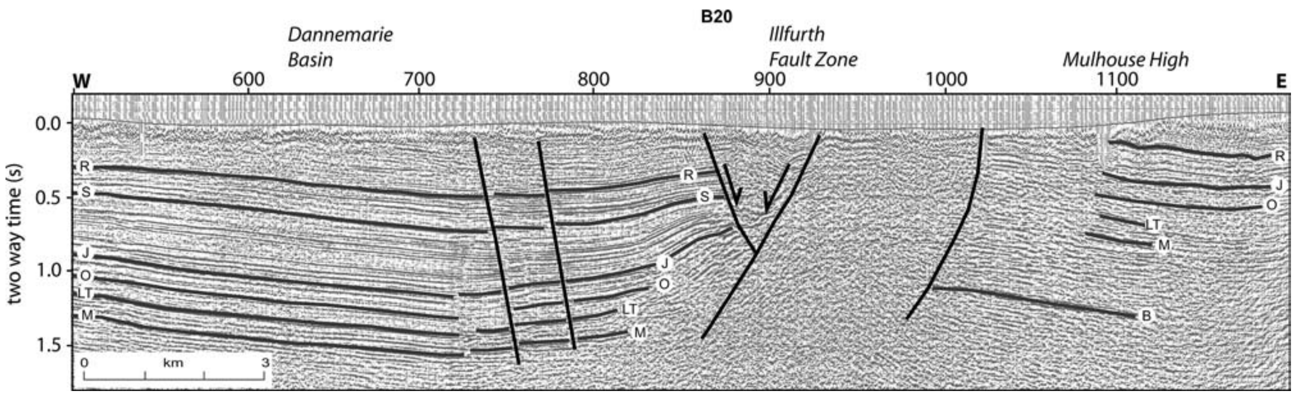


Fig. 5 Seismic section B20 showing the Illfurth Fault and the adjacent eastern part of the DB, and the Mulhouse High. For location see Fig. 2 and for markers see Fig. 4

to examine the structure of the DB during the rifting period, an isopach map of the syn-rift sediments was compiled (Fig. 6). In this map, the DB appears to exhibit the well-known NE–SW trend. In contrast, the NE–SW trend is not clearly apparent in the present shape of the basin (Fig. 7) and it becomes apparent only when its extension into the main part of the URG is also considered (e.g. Illies 1972). The two boundary faults of the basin are the Illfurth Fault (Figs. 5–7) and, on the conjugate side, the Vosges Fault that is yet to be crossed by seismic lines. Wells in the northern part of the DB include a large thickness of coarse clastic sediments, while similar sediments are not widespread further south in the basin, or close to the Illfurth Fault (Jung and Schneegans 1930; Sittler 1969; Fig. 3). In addition, the thickest accumulations of syn-rift sediments appear next to the Vosges Mountains (Fig. 6), suggesting a large boundary fault along the Vosges Mountains, rather than a gradual slope. Consequently, the DB appears to have been asymmetric in shape, the larger fault being the Vosges Fault, and a somewhat smaller conjugate Illfurth Fault (Fig. 8).

Towards the southwest end of the DB, the syn-rift sediment thickness reduces gradually, indicating that the end of the graben here is approached without a major fault. Near the Tabular Jura the data is insufficient for detailed mapping of the sediment thickness. However, some 400 m of syn-rift sediments are observed near the boundary between the DB and the Tabular Jura, indicating that the DB extended into the area that is presently the Jura. This, of course, is consistent with the small patches of rift sediments that were mapped in the Tabular Jura (e.g. Giamboni et al. 2004; Fig. 1).

The maximum throw on the Illfurth Fault was estimated to be more than 1,000 m (Nivière and Winter 2000). However, some of the displacement can be of post-rift age because post-rift tectonics is evident in the area (Giamboni et al. 2004) and may have changed the geometrical configuration of the blocks. For example, compression along parts of the Illfurth Fault is observed by the folding that appears to be associated with the fault in some parts (Fig. 5). This compression appears to be post-rifting, as it affects also the late Oligocene and younger sediments. In addition, Rotstein et al. (2005) suggested that the increased erosion of post-rift sediments on the NW part of the MH is the result of a recent uplift of this part of the MH. Based on the base Tertiary depth map (Fig. 7), the dip slip on the Illfurth Fault diminishes significantly towards the south and almost heals up when the frontal fold of the Jura is approached. Other faults in the DB that were active in the rifting period are few and mostly appear near the Illfurth Fault. Thus, it appears that the DB was relatively undisturbed internally during rifting, while most of the deformation was concentrated at its boundaries.

Post-rift structures

Following the end of the main rifting phase in middle Rupelian, the next tectonic phase was the Miocene Jura folding, forming part of the Alpine orogenic front.

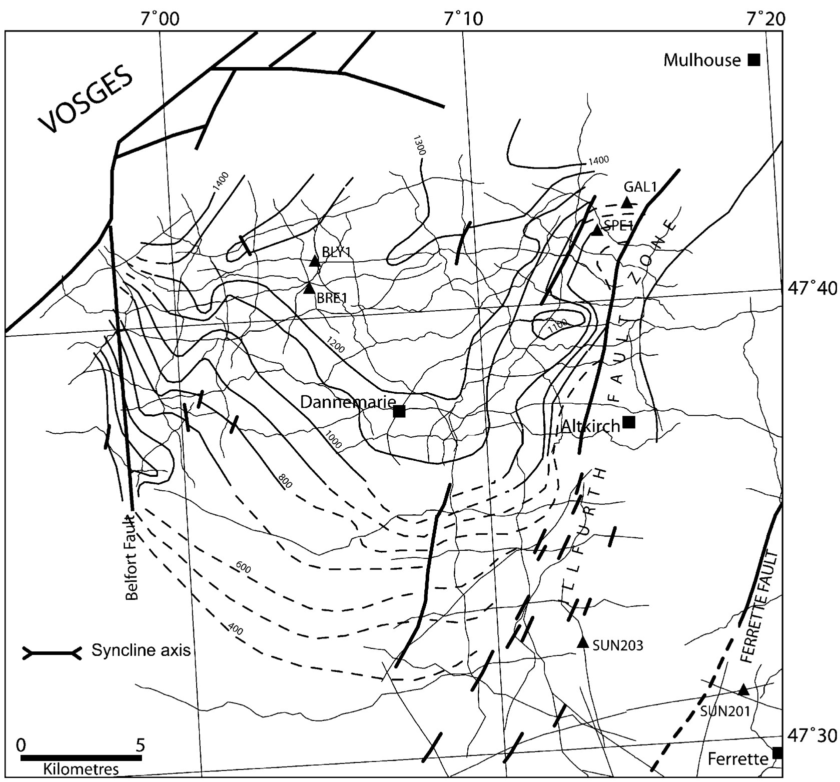
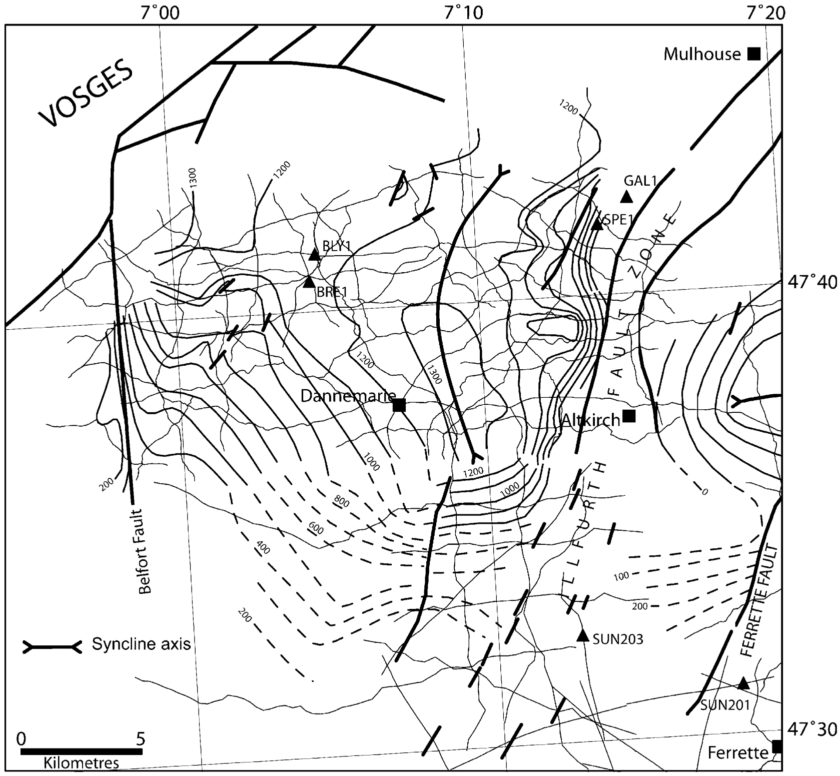


Fig. 6 Isopach map (in meters) of rift period sediments (between J and R markers), showing the shape of the DB at the end of the main rifting period

Fig. 7 Depth map of the base Tertiary unconformity (J marker), showing the present shape of the DB. Reference is sea level, with positive values depicting depths in meters below sea-level and negative values above sea-level



Several prominent Jura folds appear south of the MH, but the DB itself abuts the Tabular Jura, where folding is not developed. The depth map of the Rupelian marker (R) in the DB, associated with the end of the main rifting phase and expected to record post-rift tectonism, does not show the presence of folds (Fig. 9). There is also little evidence in the map for the NE–SW trend that characterized the DB during the syn-rift period and the post-rift basin appears to be an open syncline, quite different from the intensely folded Jura folds. It is also not spatially related to folds in the MH (Fig. 9), argued by Rotstein et al. (2005) to be buried Jura folds, and the equivalent structure across the Illfurth Fault is an anticline.

The Illfurth Fault is still apparent in the R depth map by the elevation difference of the R marker across it, particularly in the center of the DB. This elevation difference is significantly less that in the J map, or in the rift period isopach map, and there is no direct indication that the fault was active in the post-rift period. Towards the Tabular Jura, the seismic information from the DB is not complete; nevertheless, it

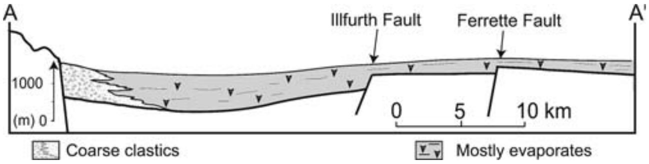


Fig. 8 Simplified cross-section of the Dannemarie basin, showing syn-rift infill pattern and the asymmetric nature of the basin

appears that the Illfurth Fault mostly healed up. The absence of marked Jura tectonics in the DB is consistent with the observation that the Tabular Jura, immediately south of the DB, is also without strong compression and intensive folding. Thus, the spillover of Jura compression into the URG is limited laterally. It occurs only in the MH, where the surface Jura folding extends all the way up to the present boundary between the two provinces.

The stress regime associated with the Jura compression phase is not yet clear. Most investigators describe the maximum principal stress direction ($\sigma 1$) to have a NW–SE trend near the URG (Bergerat 1987; Villemin and Bergerat 1987; Schumacher 2002), while others find it to be N–S oriented (Larroque and Laurent 1988). However, all these works find that during the Quaternary and possibly the Pliocene, the presently observed NW–SE compression was already active in the URG. This direction is evident by earthquake source mechanisms (Ahorner 1975; Bonjer et al. 1984; Larroque et al. 1987; Bonjer 1997; Plenefisch and Bonjer 1997) and in situ stress measurements compiled by Müller et al. (1992). As frequently noted (e.g. Ahorner 1975; Illies 1977; Villemin and Bergerat 1987; Ziegler 1994; Schumacher 2002), this stress direction is consistent with sinistral NNE–SSW strike slip motion in the URG and is likely to have activated the longitudinal faults in the graben. The timing of the onset of the main phase of Jura compression in this area was recently determined to be late Miocene (10.5 Ma, Giamboni et al. 2004). It is also not possible to determine if the onset of strike slip

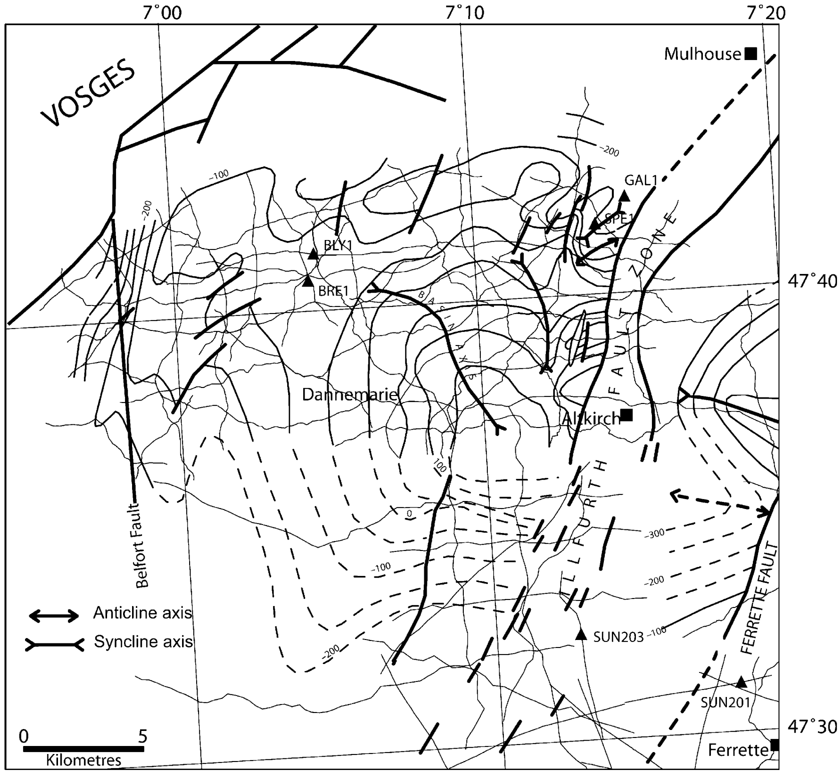


Fig. 9 Depth map of the R marker, showing the shape of the post-rift DB. Positive values are below sea-level and negative values above sea-level. Note the marked difference from the syn-rift basin (Fig. 6)

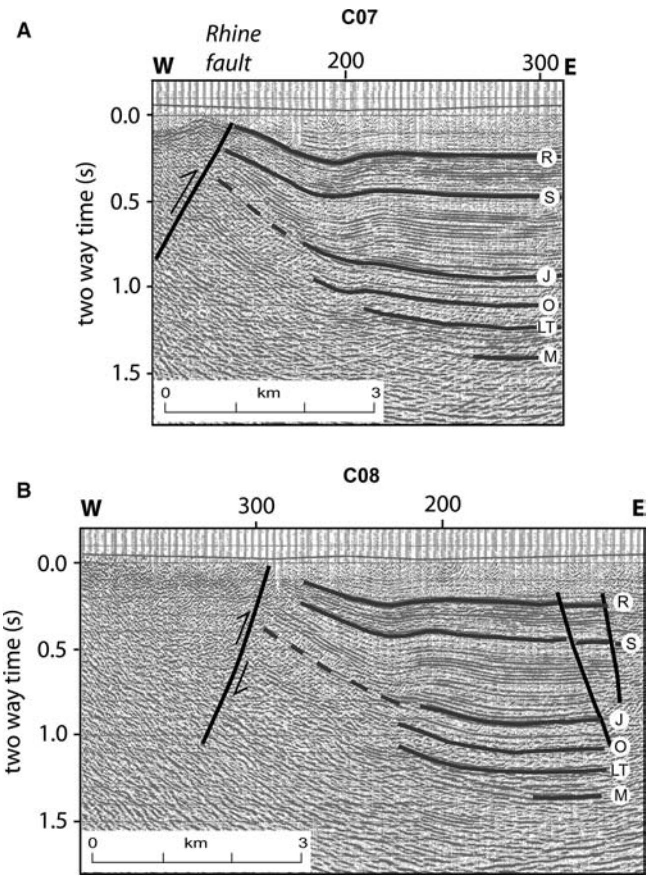


Fig. 10 Seismic sections C7 and C8 across the Belfort Fault showing the transpression, evident by the syncline in front of the fault. For location see Fig. 2 and for markers see Fig. 4

faulting in the URG was contemporaneous or post-dated the main phase of Jura folding.

We find that most of the faults inside the DB extend to the near surface. They are usually near vertical, exhibit thickness changes such that continuity can not always be reversed by a vertical shift and are occasionally associated with simple flower structures (Fig. 4). Therefore, we interpret the post-rift faulting inside the DB as having a dominant strike slip component. Most of these faults have a NE–SW trend, but a few are trending nearly N–S and a few others appear to have an ENE–WSW trend. As these faults are generally not large, only a few extend from one seismic line to the next. Without characteristic topography that is often associated with normal and thrust faulting, it is difficult to ascertain that these are indeed the same faults. Therefore, the faults that we extend between lines are at places somewhat speculative.

The SW boundary of the DB appears to be characterized by a distinct N–S trending steep gradient, associated with a known surface fault (Menillet et al. 1989), that here has been named the Belfort Fault (Fig. 9). This trend is not observed in the syn-rift isopach map (Fig. 6) and syn-rift sediments, also appear west of the Belfort Fault (Fig. 1), suggesting that the Belfort Fault is a post-rifting fault. Its linearity and trend suggests that it may accommodate the sinistral strike slip motion that is thought to prevail in the URG since the Miocene. However, at this time, it is not possible to determine more precisely the age of this fault. The trend associated with the Belfort Fault in the R depth map and at the surface (Fig. 2) results from a marked uplift of the young sediments towards the fault, rather than a dip slip on a normal fault. Several seismic lines crossing the western edge of the basin are all characterized by apparent compressional structures, and the sediments next to the Belfort Fault in each of these lines display a small syncline (Fig. 10). The sediments west of this syncline are uplifted by several hundred meters and dip to the east (Fig. 10a). The wedge shape of the syn-rift reflector package suggests normal faulting in the rifting period. However, the small and gradual thickness changes are not consistent with a larger dip of the entire uplifted sequence, which is observed in the seismic data. In addition, wherever the post-rift sediments are observed above the R marker, the uplift clearly affected them as well. Thus, the uplift postdates the deposition of the youngest observed sediments in this area. This post-rift folding along the Belfort Fault is not associated with a prominent normal fault that was active at that time. Thus, it is also not a result of forced folding, as was recently suggested for some of the folds in the URG (Bourgeois et al. 2001; Maurin and Nivière 2000); instead, it is likely to be related to a thrusting component on the Belfort Fault (Fig.10).

Discussion

The present structure of the DB is the result of the main phase of URG rifting in the Eocene-lower Oligocene, modified by subsequent changes in the stress regime. The original shape of the DB is a NE–SW elongated and slightly asymmetric graben. This trend is consistent with Variscan faults in the basement and basin geometry may be controlled by them (Edel et al. 2002; Schumacher 2002). The present shape of the basin, as observed in the base Tertiary depth map (Fig. 7), is somewhat different than the one shown previously (e.g. Illies 1972; Edel et al. 2002). It is clearly affected by the post-rift tectonism, which shifted the depocenter further to the south. The present data set resolves the change between the syn-rift and post-rift structure but can not resolve changes within the post-rift period.

The lack of Jura folds in the DB is likely to be the result of a lower compressive stress there, as compared with the MH. The shape of the Jura folds on the MH (Rotstein et al. 2005) and in the adjacent part of the Jura is more consistent with a N–S trend of the principal stress suggested by Larroque and Laurent (1988), than the present NW–SE trend. We suggest that the present trend is more likely to be the result of a Pliocene–Pleistocene change in the stress regime described by Larroque et al. (1987). As it is consistent with the strike slip displacement on the N–S and NNE–SSE faults, it is not likely to have been effective during the Miocene Jura build-up. The largest and most important transpressive fault that originated in the post-rift period in the study area is the Belfort Fault, effectively forming one of the boundary faults of the present URG. The linear topographic feature that is associated with this fault suggests that it is still one of the more active faults in the URG system.

The DB is unique in its position off the main trend of the URG. It can be regarded also as the extension of the Saône Transfer Zone, which is characterized by strike slip faulting since the late Eocene. The simple model of Lacombe et al. (1993) to explain the extension in the Saône infers that even if the DB is taken to be part of it, it is still partially affected by the main URG extension during the rifting phase. Extension in the DB may have been larger than in the rest of the Saône Transfer Zone, because of the proximity to the Rhine Graben itself, where maximum stretching was broadly directed E–W. An alternative way to explain the DB is by invoking a model of a pull-apart basin in a strike slip environment. Such a model also assumes that the DB is part of the Saône Transfer Zone and assumes a shift of the main displacement from the Vosges boundary fault to the Illfurth Fault. The mild dip of the SW boundary of the DB is quite similar to the northern boundary of the Dead Sea pull-apart graben (Manspeizer 1985). In the Dead Sea basin, the conjugate southern side is controlled by listric faults, while in the DB and its NE extension into the Rhine graben, the equivalent structure is not clear. In both models that use either transverse extension or longitudinal extension to account for the formation of a NE–SW trending DB, there is a strike slip component of motion. Unfortunately, strike slip motion is difficult to trace and there is no record of it in the southern URG during the main rift phase, except for evidence of oblique rifting (Behrmann et al. 2003). Both models may be presently considered as possible. However, the proximity of the DB to the URG, where the largest E–W, or ENE–WSW extension occurred, makes it more likely that some local modification of the model proposed by Lacombe et al. (1993) for the entire Saône accounts also for the formation of the DB.

The lack of Jura folds in the DB and in the Tabular Jura south of it is in contrast to the abundance of Jura folds in the MH and in the Jura south of it. It shows that there is a fundamental difference between the responses of the crust on the two sides of the Illfurth Fault, but the reason for this difference is not apparent. As the internal structure of the MH and the DB are similar to that of the respective areas south of them, each block should be considered together with the Jura south of it. This is particularly true since the folded Jura south of the MH and the Tabular Jura south of the DB are known to have been parts of the URG before the Jura compression. The DB, being the deepest part in the southernmost URG, was associated with the largest normal faults and with maximum extension and, most likely, with the weakest crust. Thus, it might have been the expected site for the penetration of the Jura fold system into the URG. Since the DB and the Tabular Jura are not folded, the difference in the URG response to the Jura compression is likely to be related to other differences in the local geology across the Illfurth Fault. These differences created changes in the stress regime, as was shown for the eastern

Jura and southern Black Forest and attributed to the effect of basement ramps (Becker 1989). Whatever the reason is, the Illfurth Fault appears to be an important feature from its origin in the Eocene–Oligocene, through the Jura tectonic phase and, possibly, in more recent time as a strike slip fault (Rotstein et al. 2005).

The role and origin of the post-rift syncline that is observed in the DB (Fig. 9) is not yet clear, particularly its relation to Jura compression. The syncline is nearly square and it does not extend laterally, while the Jura compression is an extensive regional phenomenon. Its southern flank is likely to be the result of the tilting caused by the known uplift of the Tabular Jura, as described for the MH (Nivière and Winter 2000). The northern flank may be part of an unrelated uplift of the southernmost Vosges. However, the unconformity that must be associated with this uplift is too shallow to be detected in the present data.

The cause of the Vosges uplift, with its maximum in its southernmost part adjacent to the DB, is not yet clear. Various hypotheses have been offered throughout the years, but none appears to be commonly accepted. Most models explain the Vosges uplift together with the similar uplift of the Black Forest Mountains, where it is also centered at the southern part of the mountain chain. The common explanation is that of a conventional rift shoulder uplift, affected additionally by a mantle diapir centered on the volcanic Kaiserstuhl (Illies 1972; Villemin et al. 1986; Villemin and Bergerat 1987). A more recent work (Gutscher 1995) describes the shortcomings of these models and suggests that the Vosges and Black Forest are part of an Alpine flexural bulge, which extends laterally further eastward. As the seismic lines in the DB did not extend into the Vosges, it is not possible to use them to examine directly the uplift mechanism. However, some indirect insight about the problem can be derived from the uplift of several hundred meters, associated with the Belfort Fault. Erosion may have reduced some of this uplift. Nevertheless, the area east of the Belfort Fault that has syn-rift sediments on the surface (Fig. 1) and used to be part of the graben, is now morphologically outside of the URG. We suggest that this uplift is the result of a thrusting component on the Belfort Fault, in addition to a dominant strike slip motion, indicating the significance of compression in this part of the URG. Compression and uplift along strike slip faults are common, particularly in restraining bends. Where the displacement on the strike slip fault is large enough, the result of a restraining bend can be an uplift of several thousands meters, as in the case of Mt Hermon in the Dead Sea Rift (Freund 1965; Quennell 1959; Garfunkel et al. 1981). In other places, small strike slip displacement may be sufficient to cause an uplift of several hundred meters in a restraining bend, as is the case in the uplift of Mt. Carmel, Israel (Rotstein et al. 1993). The Vosges Fault, the faulted southern boundary of the Vosges Mountains, is trending NE–SW, perpendicularly to the regional compressive stress. Further to the north, the western boundary fault of the URG is a longitudinal fault that is likely to have a dominant strike slip component in the present stress regime. This geometry is a classical restraining bend (Fig. 11) with compression and, most likely, thrusting across the Vosges fault. A question that is left open is whether the compression is large enough to cause the observed uplift of the Vosges Mountains. This question cannot be answered directly. However, as an uplift of several hundred meters is associated with the weakly transpressive Belfort Fault, a much larger uplift can be expected for the Vosges where the boundary fault is perpendicular to the regional maximum compressive stress.

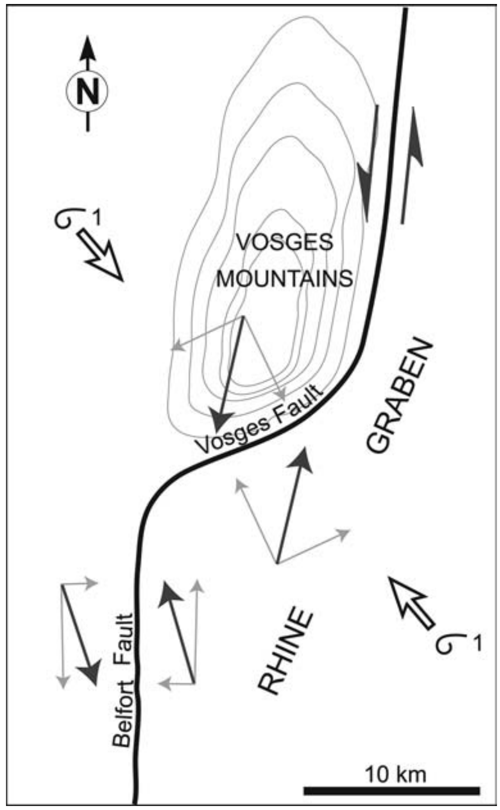


Fig. 11 A sketch showing the restraining bend on the western margin fault of the URG and the compression, which is inferred across the Vosges boundary fault. *Hollow arrows* denote the regional principal stress direction. *Other arrows* denote displacements. Where the displacement is not parallel to the main fault, it was divided into its components along and across the fault. The trend of the displacement on the Belfort Fault is deduced from the observation in the seismic data of compression across it

Using this model to explain the Vosges uplift infers that the uplift mechanism is local and that the Vosges uplift is not directly related to the uplift of the conjugate Black Forest. A similar local mechanism, related to the local response to the regional stress regime, must be looked for in the Black Forest Mountains. Although this model is somewhat speculative, it has several merits. It is consistent with the observation that the current uplift of the Vosges, may have started as late as the Pliocene (Illies 1977), or even more recently (Duringer 2001). Thus, it is approximately contemporaneous with the onset of the present stress regime and the strike slip faulting. In addition, it is consistent with the observation that the Heidelberg basin in the north of the URG is the site of Pliocene to recent subsidence of some 1,000 m (e.g. Sittler 1969). This phenomenon is attributed to the component of extension that results from the NNW–SSE trend of this part of the URG, in a left lateral strike slip environment (Illies 1978; Schumacher 2002). By the same reasoning the boundary fault south of the Vosges, almost perpendicular to the Heidelberg boundary faults, must be associated with compression, thrusting and uplift of the same order of magnitude. Finally, it is consistent with the Pliocene to recent compression in the southernmost URG, that is described by Giamboni (2004) and the uplift inferred by the absence, in this area, of the Quaternary deposits that are present further north (Schumacher 2002).

Conclusions

The present shape of the DB is the result of its formation as a NE–SW trending graben during Eocene–early Rupelian, modified by subsequent tectonism. Originally, the DB extended into the area of the present Tabular Jura and its SW boundary was not faulted significantly. Weak Jura compression affected the DB during the Miocene. It did not result in folding, but the uplift of the Tabular Jura eroded much of the syn-rift sediments in that area, and resulted in a northward dip of the sediments in the southern DB. NW compression was associated in the DB with the strike slip faulting on existing faults, as in the URG in general. It also resulted in the formation of new longitudinal faults, most notably the Belfort Fault, which can be regarded as one of the boundary faults of the present URG system. A component of thrusting on the Belfort Fault resulted in several hundred meters of uplift on the hanging-wall. Large scale transpressive uplift of the southern Vosges mountains is likely due to a restraining bend in the western boundary fault of the URG system.

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