water into steam. This phase transformation requires the absorption of the latent heat of vaporization and results in cooling of the surrounding fluid. As noted immediately above, cooling leads to upward displacement of the phase boundary and stabilization of the system. The cooling produced by the effects of advection and latent heat absorption promotes one destabilizing tendency associated with the thermal contraction of the water—colder water weighs more, and this additional weight of the column tends to drive the system in the same direction as the original downward perturbation.

To assess these effects quantitatively we have analysed the gravitational stability of horizontally infinite superposed fluid layers separated by a phase-change interface in a porous medium. The basic state consists of a motionless layer of water above a static layer of steam. Pressure increases with depth in the undisturbed state according to the hydrostatic law; the liquid and vapour densities govern the slopes of the pressure–depth profiles in the water and steam layers, respectively. Temperature is assumed to increase linearly with depth in the basic state. The stability of this configuration is governed by Darcy's law, which relates velocity and pressure perturbations, the incompressible continuity equation (density changes within each phase are negligible), which relates horizontal and vertical velocity perturbations, and the phase equilibrium equation, which relates pressure perturbations to the distortion of the water–steam interface. The differential equations for the disturbances are linearized and solved subject to the following conditions: (1) isothermal upper and lower boundaries, (2) no flow through the lower boundary, (3) constant pressure on the upper boundary, (4) continuity of pressure, temperature and mass flow at the water–steam interface, and (5) balance of conductive heat flux and latent heat release at the phase-change boundary.

The analysis shows that stabilization of water overlying steam requires

$$K<\frac{2k}{L\left(\frac{\mathrm{d}p}{\mathrm{d}T}\right)_{\mathrm{c}}}\,\frac{\left(\frac{\bar{\mu}_{\mathrm{l}}}{\bar{\rho}_{\mathrm{l}}}+\frac{\bar{\mu}_{\mathrm{v}}}{\bar{\rho}_{\mathrm{v}}}\right)}{\left(1-\frac{\bar{\rho}_{\mathrm{v}}}{\bar{\rho}_{\mathrm{l}}}\right)}\left\{\frac{\Gamma\left(\frac{\mathrm{d}p}{\mathrm{d}T}\right)_{\mathrm{c}}}{g\bar{\rho}_{\mathrm{l}}}-\frac{\left(\frac{\bar{\rho}_{\mathrm{v}}}{\bar{\rho}_{\mathrm{l}}}+\frac{\bar{\mu}_{\mathrm{v}}}{\bar{\rho}_{\mathrm{v}}}\frac{\bar{\rho}_{\mathrm{l}}}{\bar{\mu}_{\mathrm{l}}}\right)}{\left(1+\frac{\bar{\mu}_{\mathrm{v}}}{\bar{\rho}_{\mathrm{v}}}\frac{\bar{\rho}_{\mathrm{l}}}{\bar{\mu}_{\mathrm{l}}}\right)}\right\}, \quad (1)$$

where ρ is density, μ is dynamic viscosity, $(\mathrm{d}p/\mathrm{d}T)_{\mathrm{c}}$ is the slope of the Clapeyron curve, L is the latent heat, Γ is the geothermal gradient, g is the acceleration of gravity, k is the combined thermal conductivity of the fluid and matrix (dominated by the conductivity of the rocks), K is the permeability, and subscripts l and v refer to liquid and vapour, respectively. All the quantities in equation (1) are to be evaluated at the quiescent position of the water–steam interface. Condition (1) is necessary for stability. Although it is not sufficient, an exact analysis[8] shows that the value of permeability given by the right side of equation (1) is a good estimate of the maximum permeability for which stabilization is possible.

The evaluation of equation (1) leaves little uncertainty in the largest value of permeability allowing stabilization of the condensate layer. This is because all the quantities in equation (1), except for k and Γ, depend only on temperature, and there is little temperature variation within and among the steam reservoirs of natural vapour-dominated systems. In addition, the condensate layers are generally several hundred metres thick, precluding any large uncertainty in the value of Γ. Finally, the thermal conductivities of reservoir rocks are known to within factors of 2 or 3. Using a value of 242 °C for the temperature at the condensate layer–steam zone interface (see Fig. 1), $\Gamma = 0.555\ °\mathrm{C\,m^{-1}}$ and $k = 4\ \mathrm{J\,m^{-1}\,s^{-1}\,K^{-1}}$, we find that K must be less than 80 nm^2 (0.08 millidarcy) for the condensate layer to be gravitationally stable [a more exact analysis[8] shows that $K \lesssim$ 40 nm^2 (0.04 millidarcy) for stability].

A permeability of 40 nm^2 is more than an order of magnitude smaller than permeabilities generally thought to occur in the main reservoirs of geothermal systems. It is not surprising, however, that relatively low permeability is required near the condensate layer–steam zone interface to keep the water from falling down into the underlying region of steam. The significance of our result is the fact that any stabilization is possible for a non-zero value of permeability. The natural discharge of vapour-dominated geothermal fields attests to the through-flow of steam within the system. This could not occur if some impermeable layer were required to seal or cap the main steam reservoir to stabilize the condensate layer. It is the special behaviour of the phase-change interface between water and steam that permits stabilization of the systems with a non-zero permeability; if the interface behaved in a manner similar to the boundary between immiscible fluids the rocks would have to be impermeable to keep the condensate layer from falling down.

The low permeability required for the stability of the condensate layer probably determines the thickness of the layer. During the formation of a wet geothermal system, rising steam from the relatively permeable main reservoir penetrates rocks whose permeability decrease with proximity to the surface. The top of the main reservoir will extend upwards until condensate can be stabilized by rocks of sufficiently small permeability.

We thank Geothermal Energy N.Z. Ltd, managers of the Kamojang Project, acting on behalf of the Indonesian and New Zealand governments, for supplying the data used in Figs 1 and 2. This work was supported, in part, by NSF grant ENG 76-82119.

Received 28 April; accepted 22 July 1980.

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The Gulf of Lion: subsidence of a young continental margin

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The origin of the subsidence of Atlantic-type continental margins is of much interest to geology and geophysics. The main problem at these margins is in explaining the substantial thicknesses of mainly shallow-water sediments which accumulate soon after rifting and plate separation. We have now used biostratigraphic data from commercial wells to study the subsidence history of a young Atlantic-type continental margin. Backstripping the sediment load reveals that the margin has been subsiding at nearly mid-ocean ridge rates. Geological data indicate that lithospheric stretching alone cannot account for this amount of subsidence and other processes must be involved.

Studies of biostratigraphic data from deep commercial boreholes suggest that the tectonic subsidence of margins is thermal in origin[1–5]. Sleep[1] suggested the subsidence is due to cooling of the lithosphere following uplift and erosion. This model cannot, however, fully explain the amount of tectonic subsidence observed at these margins[3] or the lack of evidence of domal uplift during rifting[6]. McKenzie[7] proposed a model in which the subsidence is due to cooling of the lithosphere following stretching and thinning at the time of rifting. This model seems to explain the tilted and rotated blocks bounded by listric faults which develop during rifting[8,9].

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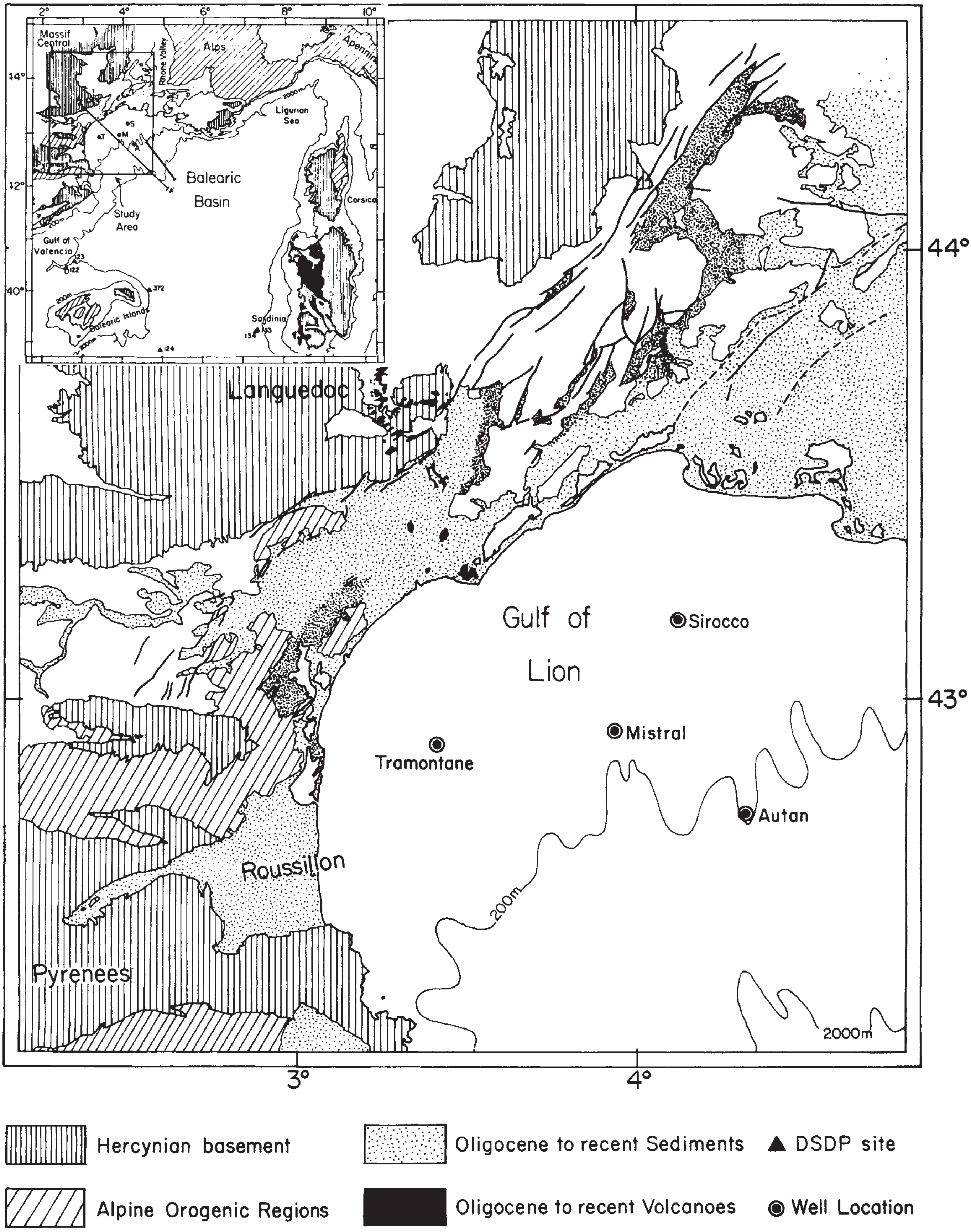


Fig. 1 Geology of the Gulf of Lion. The heavy lines are faults which have been active during the Oligocene, dashed where inferred. The heavier dots show the location of Oligocene red beds. Inset, generalized geology of the regions surrounding the Balearic Basin and location of Gulf of Lion. The heavy line is the location of the 'Christiane' multichannel seismic profile[30,32], A–A' is the location of the profile shown in Fig. 2.

By quantitatively accounting for the effects of sediment and water loading the tectonic subsidence of a margin can be isolated[2–4]. To correct for sediment and water loading, the sedimentary section should be reconstructed at intervals during the development of the margin, taking into account the effects of compaction, changes in water depths and eustatic sea-level changes. The sedimentary section should then be isostatically unloaded, or backstripped[2,3].

Most previous studies have been based on stratigraphic data from the thickly sedimented, relatively old margins, such as eastern North America. The wells in these studies generally did not sample the pre-rift and syn-rift sediments or crystalline basement rocks and therefore do not provide either a complete record of the subsidence history or a reliable age for the opening of the central North Atlantic. A detailed knowledge of the early subsidence history is important in discriminating between different models for the origin of the margin.

A good example of a well sedimented young Atlantic-type continental margin is the Gulf of Lion in the Balearic Basin (Fig. 1). Geophysical data indicate that the Balearic Basin contains up to 6 km of sediments, including over 1 km of Messinian evaporites, overlying a thin oceanic-type crust[10–12]. Geological and palaeomagnetic evidence[11,13–16] suggests the basin formed by Oligocene/Miocene rifting of Corsica–Sardinia from France (Fig. 1). Messinian evaporites had been interpreted[17] as indicating that the western Mediterranean was not a deep basin until post-Miocene time. However, DSDP site 372 east of Menorca[18] established that the western Mediterranean was already a deep basin by the Lower Burdigalian (~20 Myr BP). Although the tectonic setting of the basin is poorly known, it may have formed as a back-arc basin behind a migrating island arc–trench system to the east[16]. Extensive Oligocene through Lower Miocene andestic volcanism related to this arc is found in western Sardinia[19].

Figure 1 summarizes the geological setting of the Gulf of Lion. The Eocene in southern France was a period of overthrusting associated with the Pyrenean orogeny[20,21]. The Oligocene, in contrast, was a period of extension and normal faulting[20,21], much of which seems to have reactivated older fault trends. Continental sediments accumulated in NE–SW trending grabens in Languedoc and Roussillon. Similar extension and graben structures also occurred in Sardinia and along the margins of the Balearic Islands[22,23]. Drainage patterns in southern France during the Eocene and Oligocene were towards the north and west, away from the present Mediterranean[24,25]. Then, in the Lower Miocene (Aquitanian) a major transgression occurred marking the opening of the Balearic Basin[26,27].

The structure and stratigraphy of the Gulf of Lion margin is known from commercial drilling and seismic reflection data. Three of the wells drilled recovered over 3 km of Miocene to Recent sediments overlying Palaeozoic basement[28,29]. In addition, the Autan well (Fig. 1) recovered 244 m of ?Oligocene red beds[29]. The seismic data in the Gulf of Lion[30–33] show a gently dipping Pliocene to Recent sedimentary section unconformably overlying a more steeply dipping Miocene section. Information on the basement structure of the Gulf of Lion is sparse but available data[28,33] indicate the Miocene to Recent sedimentary section overlies horst and graben structures which are infilled by continental sediments of probable Oligocene age (Fig. 2). Thus, the available geological and geophysical data show that the structural evolution of the Gulf of Lion region margin was similar to that of a normal Atlantic-type continental margin[12].

We have used biostratigraphic data from three wells[29] in the Gulf of Lion (Figs 1, 2). The sediments in each well were backstripped[2,3] to obtain the tectonic subsidence of the margin. The wells were corrected for compaction using available sonic, density and porosity logs (L. Montadert, personal communication) and backstripped with the Airy model of isostasy[2,34] and a mantle density of 3.33 g cm^{-3}. The palaeobathymetry for the wells was based on the micropalaeontological studies of Cravatte *et al.*[29]. We did not correct for eustatic sea-level changes as they are not well known in the Neogene due to uncertainties in the timing of Antarctic glaciation.

The sedimentary record in the wells is interrupted by a major erosional unconformity due to the Messinian salinity crisis[30,35] (Fig. 3). The unconformity is observed as a discontinuity in the lithological logs. We backstripped each well using as wide a range of estimates for the magnitude of the erosional event as

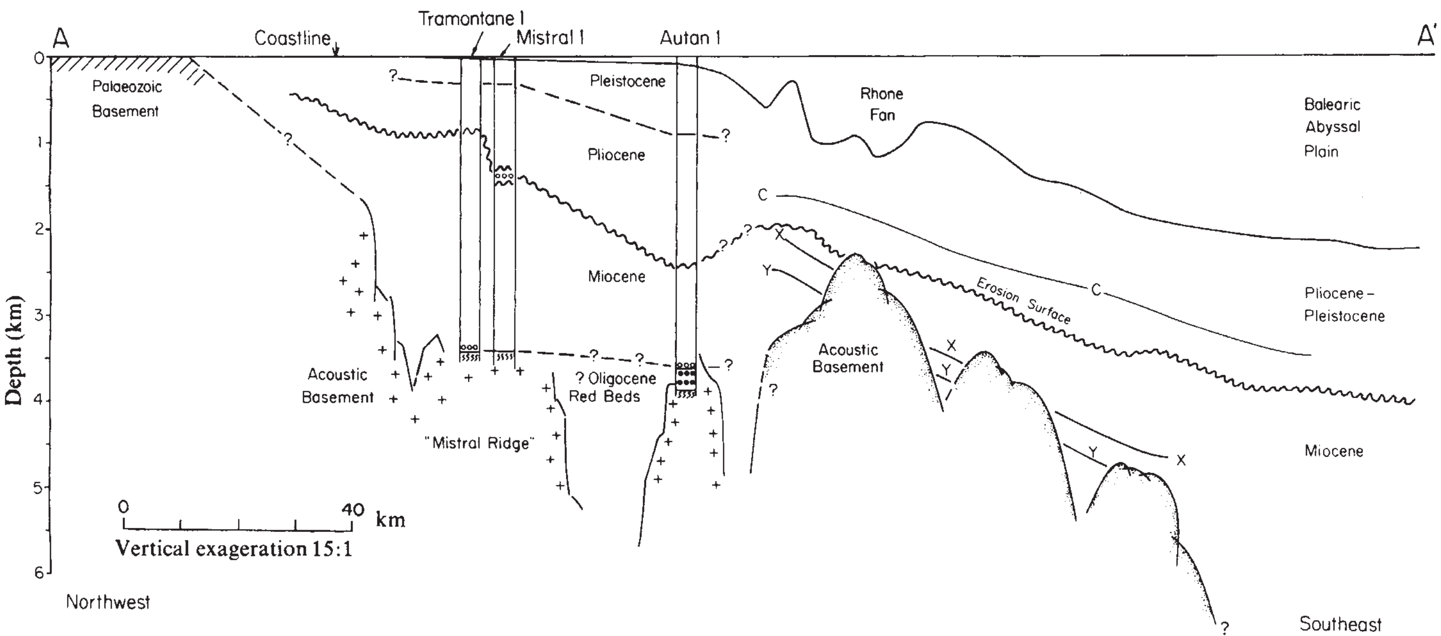


Fig. 2 Profile across the Gulf of Lion margin. Prominent seismic reflectors and acoustic basement are projected from the 'Christiane' profile[30]. In the wells: ○, basal conglomerates; ●, continental red beds. The wells bottomed in Palaeozoic schists[28,29]. The basement structure suggests prominent horsts and grabens infilled by red beds (refs 28, 33 and L. Montadert, personal communication) seaward of a hinge zone.

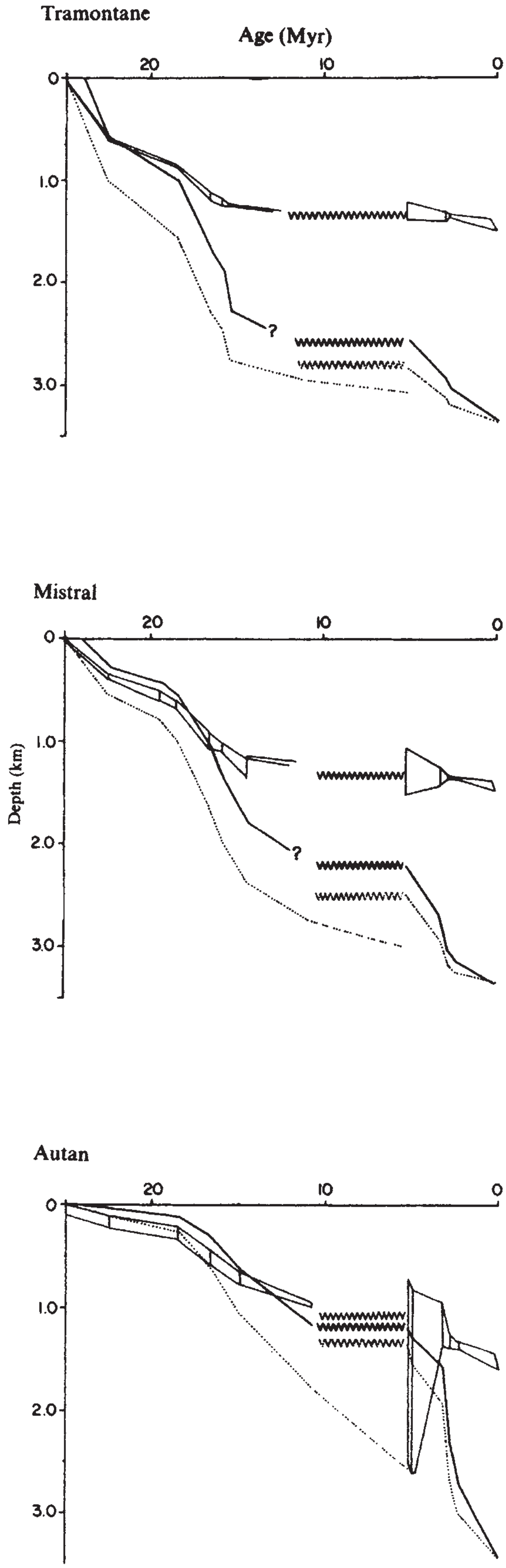


Fig. 3 Subsidence in the three wells studied. The heavy solid lines are the observed sediment thicknesses. The dotted lines represent the reconstructed sediment accumulation after correction for compaction, dashed where later removed by erosion. The double solid line is the subsidence after backstripping. The width of the lines represent the range of estimates for the water depth based on micropalaeontological studies[29].

was compatible with the well logs and found that in each case the trend of the tectonic subsidence (Figs 3, 4) was continuous through the salinity crisis. This agrees with the interpretation of the Messinian event as an evaporitic drawdown[32,35].

Using the observation that the tectonic subsidence was continuous, it is possible to estimate the amount of erosion of the margin during the salinity crisis. We found that 1,200–1,500 m of sediments were eroded at Autan and that much less were eroded at the more landward Mistral (380–600 m) and Tramontane (<320 m) wells. The water depths immediately following the salinity crisis were estimated to be 700–900 m at Autan and correspondingly smaller at the more landward wells, in agreement with those determined from micropalaeontological studies[29]. At the Autan well, however, by using the trend of the tectonic subsidence we can estimate a much smaller range of water depths than that which can be estimated from micropalaeontology (bathyl).

The tectonic subsidence at each well (Fig. 3) constitutes less than half of the total sediment accumulation and continues smoothly through the Messinian event. The tectonic subsidence is similar in form for each of the wells. The main difference occurs in the early subsidence which is larger in a landward direction. Small variations in the rate of the tectonic subsidence appear to correlate from well to well and may be due either to changes in sea-level[36] or to uncertainties in absolute age dating of the sediments.

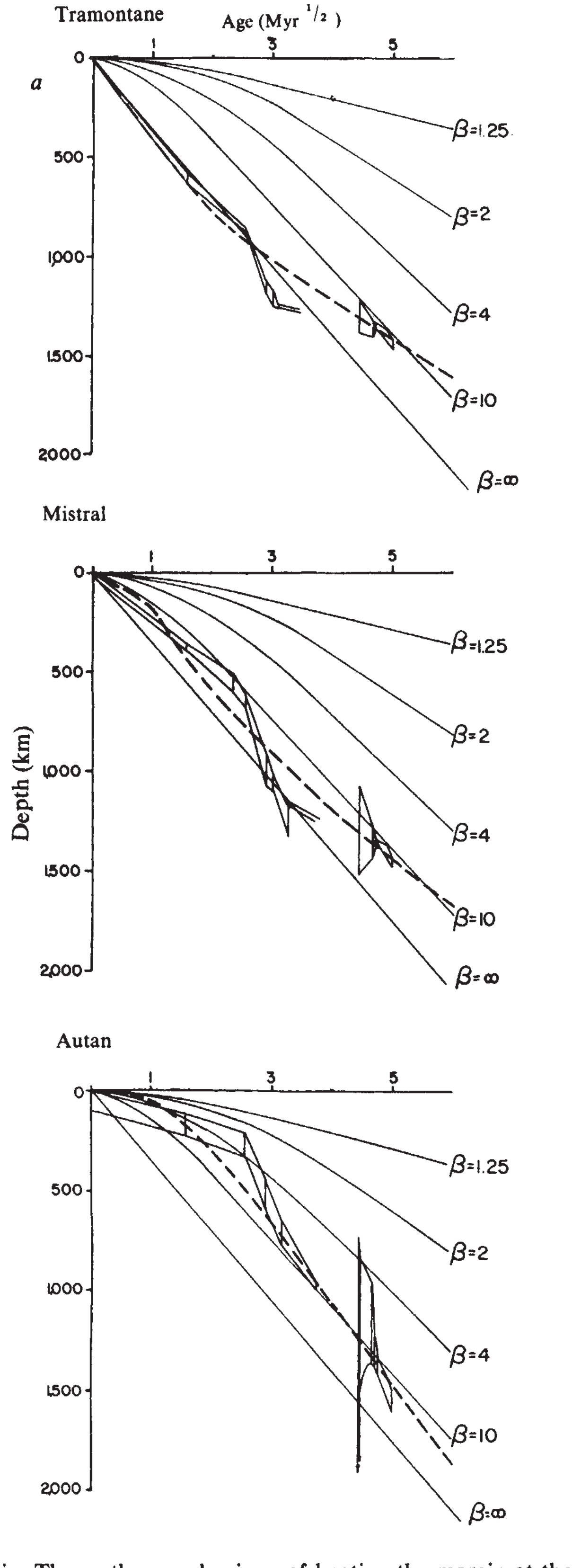
The most likely cause of the tectonic subsidence is thermal contraction following stretching and thinning of the margin at the time of rifting. A useful way to examine the thermal subsidence of a margin is to plot the tectonic subsidence against $\sqrt{\text{age}}$ (ref. 7). Figure 4 shows the tectonic subsidence for the wells along with that predicted by a lithospheric stretching model[7]. The origin in Fig. 4 is 25 Myr BP which is the best estimate for the age of opening of the basin[13,14,16,27]. The parameter $\beta = \infty$ is equivalent to the subsidence of a mid-ocean ridge.

The total tectonic subsidence at each well is large and seems to exceed $\beta = 10$. The fit of the subsidence data to the model, however, is poor. The tectonic subsidence early in margin history is progressively greater in a more landward direction and at Tramontane exceeds that of a mid-ocean ridge.

Numerical modelling of the continental margin off Norway[37] has shown that lateral transport of heat towards the continent may be important in the early history of the margin. We have, therefore, modified McKenzie's[7] model to include a finite crustal thickness for oceanic lithosphere and the effects of lateral as well as vertical conduction of heat. The subsidence was computed for a two-dimensional model in which β varied across the margin (Fig. 4*b*). The fit to the subsidence data has been greatly improved (Fig. 4*a*) and the estimates of β (2.6 to 6) at the wells reduced. These considerations suggest that thermal cooling, both vertical and lateral, of a hot, thinned lithosphere can explain the tectonic subsidence of the margin.

The magnitude of the tectonic subsidence implies that the continental basement in the Gulf of Lion was greatly extended during rifting. Extension of this magnitude, however, does not seem to be compatible with the geology of the Gulf of Lion. Normal faulting and graben structures containing Oligocene red beds occur both on land[20,21] and off-shore (ref. 33 and L. Montadert, personal communication). However, the volume of these pre-rift and syn-rift sediments is relatively small compared with the volume of the post-rift sediments (Figs 1, 2). All the wells drilled in the Gulf of Lion (Fig. 1) contain little or no Oligocene sediments (Fig. 2) and are located on horsts. If the lithosphere had been stretched by the estimated values of β then the sediment accumulation during rifting should have been in excess of 4 km. Thus, the small observed subsidence associated with rifting precludes large amounts of stretching of continental crust, while the magnitude of the thermal subsidence requires extensive heating of the lithosphere during rifting.

These results suggest that although there was extension during the Oligocene, the passive heating resulting from stretching[7] cannot fully account for the observed tectonic subsidence of the



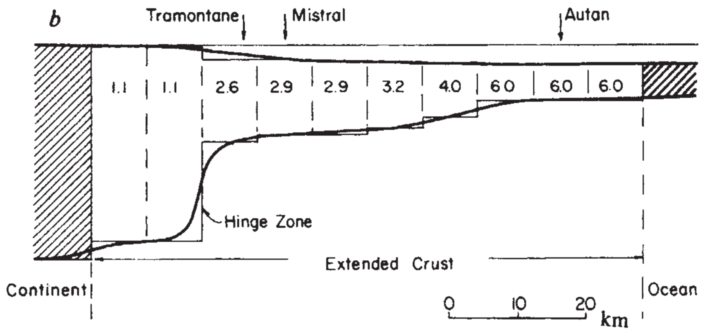


Fig. 4 *a*, Tectonic subsidence plotted against square root of the age. Width of lines represents range of estimates for water depth[29]. Age of opening was assumed to be at 25 Myr BP. Light lines, theoretical subsidence calculations from lithospheric stretching model[7]. $\beta = \infty$ corresponds to mid-ocean ridge subsidence. Dashed lines, subsidence calculations for two-dimensional model illustrated in *b*, Crustal section of the two-dimensional model used to estimate subsidence at the wells. Dashed lines represent boundaries of the blocks used for calculations. The values for each block are the estimates of β. The heavy lines illustrate the crustal thinning that occurred at the margin at the time of stretching (25 Myr BP). Note the large values of initial water depths. No vertical exaggeration.

margin. Thus, other mechanisms of heating the margin at the time of rifting are required. One possibility is that a thick crust and relatively high thermal gradients already existed in the lithosphere before rifting, due to Pyrenean Eocene overthrusting and/or andesitic volcanism in the nearby island arc–trench system. Alternatively, continental rifting may be caused by deep seated processses in the mantle which result in an active heating of the lithosphere at the margin. Further work on other Atlantic-type continental margins is needed to distinguish between these possibilities.

We thank L. Montadert for the sonic, density and porosity logs for the wells. This work was supported by a Phillips Petroleum Company Fellowship to M.S.S. and NSF grant OCE 77-10647. Lamont–Doherty Contribution no. 3036.

Received 14 April; accepted 4 July 1980.

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