Physical explanation of the formation and evolution of inversion zones and marginal troughs

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ABSTRACT

Inversion zones are elongate structures, some tens of kilometers wide and up to hundreds of kilometers long, that have deformed in response to compression and produced topography. Inversion zones in the Alpine foreland are mainly associated with Mesozoic grabens and troughs, and although very important in the geologic picture, the conditions of their formation and evolution and their regional geologic significance are not entirely understood. The internal structure of inversion zones is variable and depends on details in the pre-inversion setting, the inversion-inducing stress field, and the sedimentary fill. However, on a larger scale, most inversion zones share certain principal observational features, which sample the physical structure and the rheologic properties of the lithosphere and thereby provide an opportunity to test hypotheses of lithospheric rheology and dynamics. The quantitative model presented in this paper explains how inversion zones and the associated marginal troughs are related to lithospheric zones of differential shortening and regional isostatic compensation of the induced topography.

Keywords: basin inversion, numerical modeling, rheology, continental lithosphere.

INTRODUCTION

The term "inversion" refers to the reversal of displacements on faults. However, more recently "inversion" has become almost synonymous with specific tectonic processes in which the deeper parts of a sedimentary basin or graben become uplifted and eroded. This evolution in terminology is evidenced by the emergence of the phrase "sedimentary basin inversion."

Some inversion structures in the Alpine foreland are related to transpressional uplifting of major basement blocks along old fracture lines as, for example, the Lusatian block in the Bohemian massif, whereas others developed from wrench induced Mesozoic grabens and troughs (Ziegler, 1987; Ziegler et al., 1995). The development of marginal sedimentary troughs contemporaneously with the inversion are characteristic features of inversion zones. The Late Cretaceous and Paleocene inversion structures of the Alpine foreland often are associated with marginal chalk depocenters, which exhibit some degree of symmetry around the inversion axis and which are deepest closest to the inverted zone and shallowest away from it (Figs. 1 and 2).

The time between formation of a graben or trough in northwest Europe and its inversion is largely bracketed between the general phases of rifting and wrench faulting relating to the Mesozoic breakup of Pangea and the phases of Late Cretaceous and Cenozoic compression due to the collision of the African and Eurasian plates. The formation of a basin in transtension brings the mantle closer to the surface, and as the stretching-related thermal transients dissipate, the mantle becomes relatively cold and strong. From this consideration emerges the enigma (Ziegler et al, 1995) of basin inversion: Why does transpressional inversion take place in the central part of a basin where the upper mantle is relatively strong when the flanking structural highs have a deeper and warmer and therefore weaker mantle? The model presented here offers a solution to this problem.

On the basis of a one-dimensional study, Sandiford (1999) has argued that an extensional sedimentary basin in thermal equilibrium may have the highest upper mantle temperature and therefore the weakest lithosphere where the basin is deepest. However, when considering the effect of heat refraction, we find that (1) the maximum Moho temperatures almost always occur beneath the flanking basement highs of an extensional basin, and (2) it requires extreme sedimentary structures for the Moho temperature to be at maximum at the center of the basin.

Therefore, we do not favor thermal weakening as the vehicle of transpressional basin inversion. Rather, together with others (Ziegler, 1987; van Wees and Stephenson, 1995; Ziegler et al., 1995), we invoke the existence of a zone of structural and rheologic weakness, which is utilized in transtensional basin formation and later reused in transpressional inversion. This hypothesis also applies to the inversion of major basement blocks, which, like the Lusatian block in the Bohemian massif (Fig. 1), have remained positive without sedimentary cover since Paleozoic time and yet have become deformed in transpressional inversion and produced a marginal trough (Malkovský, 1987). We have investigated the consequences of the presence of a zone of weakness by performing numerical experiments on a

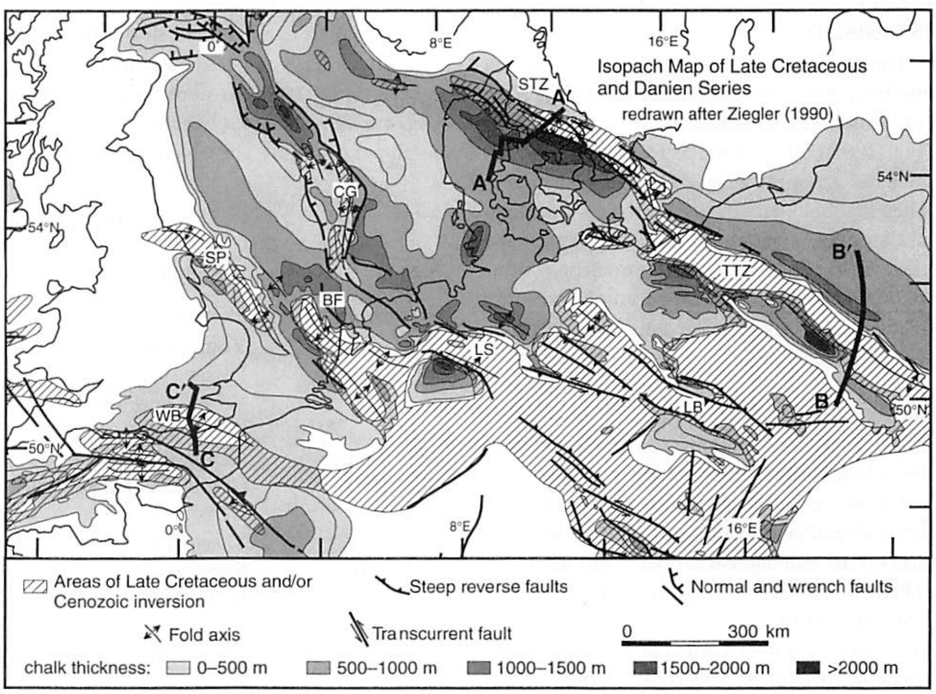

Figure 1. Chalk isopach of northwest Europe (Ziegler, 1990). Areas of Late Cretaceous and/or Paleocene inversion are flanked by characteristic chalk depocenters. BF—Broad Fourteens, CG—Danish Central graben, LB—Lusatian block, LS—Lower Saxony basin, SP—Sole Pit basin, STZ—Sorgenfrei-Tornquist zone, TTZ—Tornquist-Teisseyre zone, WB—Weald basin. A–A′, B–B′, and C–C′ are locations of cross sections in Figure 2.

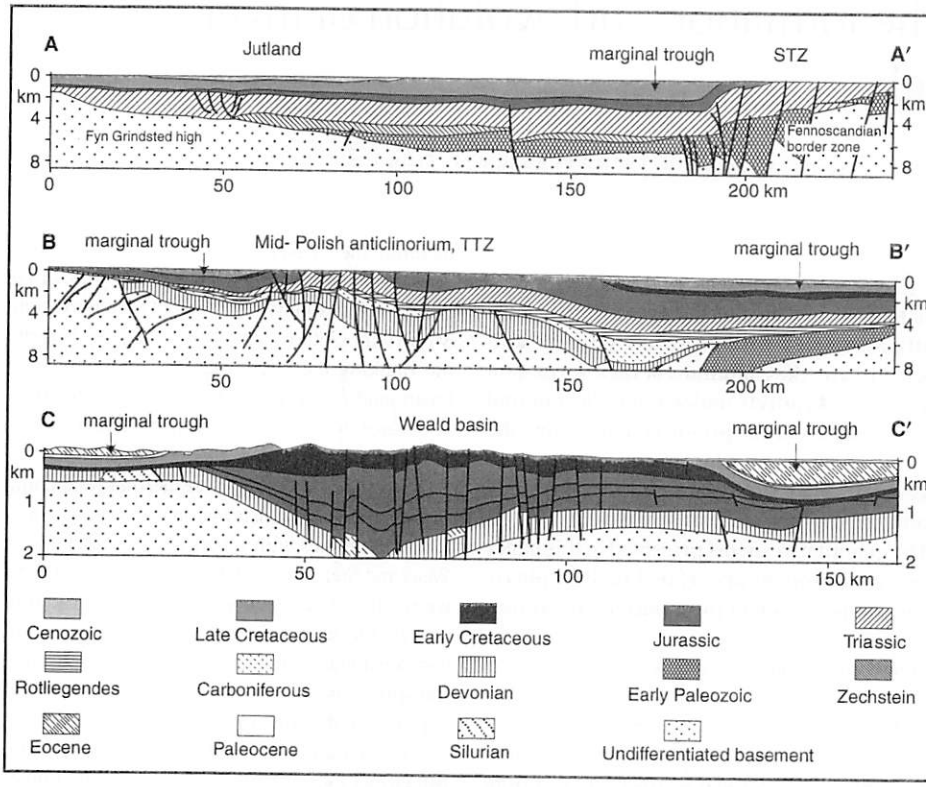

Figure 2. Inversion profiles. A–A′—Sorgenfrei-Tornquist zone (STZ; Ziegler, 1990), B–B′—Tornquist-Tesseyre zone (TTZ; Ziegler, 1990), C–C′—Weald basin (Chesher, 1991).

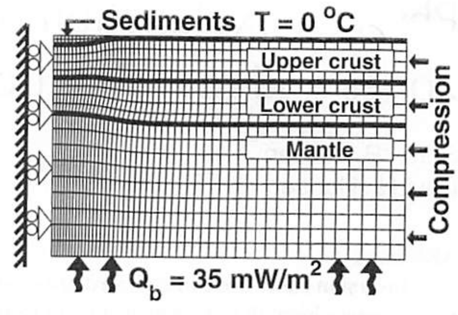

Figure 3. Finite element model of lithosphere. Each square is composed of four triangles. Initial total crustal thickness is 40 km. Model parameter values are given in Table 1.

rheologically layered lithosphere model and comparing the results to observations.

FINITE-ELEMENT MODEL OF INVERSION

The numerical-model concept follows Braun and Beaumont (1987) and Hansen et al. (2000). It is a thermomechanical finite-element model (Fig. 3) that allows for elastic, plastic, and viscous deformation. Plastic deformation occurs when the yield strength of the rock is exceeded and simulates fracturing. The location of the yield surface in stress space depends on confining pressure, rock type, temperature, and strain rate. Buoyancy forces act at density-contrast surfaces located at the surface, middle crust, base of the crust, and base of the lithosphere. The loading effects of sedimentation and erosion are treated by surface tractions. Isochron markers in the sediments trace the structural evolution. The sediments compact irreversibly with depth. The thermal feedback from sedimentation and erosion on the mechanical properties of the lithosphere is considered. The left and right vertical boundaries of the model are boundaries of symmetry. Thus, numerical experiments result in periodic structures. The right vertical model boundary is displaced laterally with no tilt at a constant velocity to produce compression or extension. The resulting force is the force needed to deform the lithosphere for the imposed lateral displacement rate. The thermal boundary conditions are a constant surface temperature, a constant basal heat flux, and no lateral flux across the vertical model boundaries. Model parameters are given in Table 1.

The weak zone is formed in the numerical model by a Gaussian-shaped reduction of the yield strength, T_0, in the upper crust and of the creep activation energy, Q, in the lower crust and the upper mantle. The maximum reduction amounts to 80% in T_0 and 10% in Q in the center of the structure. The reduction in T_0 makes plastic flow occur at smaller deviatoric stress levels and simulates the presence of many isotropically distributed and easily reactivated fault planes or some other continuous mode of strength reduction. The reduction in Q reduces viscosity. The changes in T_0 and Q can be considered as empirical means necessary for the model to yield inversion structures that resemble the reality of intense deformation in the central inversion zone and little deformation in the adjacent sedimentary troughs. The exact mechanisms of mechanical weakening are not the subject here. Rather, we investigate the consequences of the presence of a weak zone. In some cases, for example, the Weald basin (Figs. 1 and 2C), low-angle crustal faults accommodate the majority of crustal shortening (D. Blundell, 2000, personal commun.; Lake and Karner, 1987). This situation is an example of discrete zones of intense deformation that control the formation of the central inversion structure; however, the large-scale effects of the inversion remain the same: uplift and erosion of a

TABLE 1. MODEL PARAMETERS

Symbol	Meaning	Mantle	Lower crust	Upper crust	Sediment matrix
E (Pa)	Young's modulus	10^{11}	10^{11}	10^{11}	10^{11}
ν	Poisson's ratio	0.25	0.25	0.25	0.25
n	Creep parameter	4.48	3.20	3.10	3.10
B (MPa $s^{1/n}$)	Creep parameter	0.2628	12.28	208	208
Q (kJ/mole)	Activation energy	498	239	135	135
T_0 (MPa)	Compressional strength	52.4	26.2	26.2	17.5
ρ (kg)	Density	3300	2900	2700	2700
k (W/m/K)	Thermal conductivity	4	2.3	3	2
c (J/kg/K)	Specific heat	1000	900	900	900
A (μW/m^3)	Heat production rate	0	0.25	1.5	0.75
Φ_0	Surface porosity				0.60
L (m)	Porosity decay length				2000

Note: Creep parameters are from Chopra and Paterson (1981) for mantle (wet dunite), from Shelton and Tullis (1981) for lower crust (anorthosite), and from Paterson and Luan (1990) for upper crust (wet quartzite). See Braun and Beaumont (1987) for explanation of governing thermomechanical equations. Sediment porosity decreases with depth according to $\Phi = \Phi_0 \exp(-z/L)$, where z is depth below sediment surface.

narrow and elongated central zone and the simultaneous formation of marginal troughs.

MODELING RESULTS

In the initial graben (Fig. 4, model time 0 m.y.) the pre-, syn-, and postrift sediments formed by extension of a lithosphere equipped with a preexisting crustal weak zone of effectively 60 km width. The strength of the mantle is not modified. The initial lithosphere thickness is 100 km, 40 km of which is crust. The short term response of this structure to compression does in small details depend on the deviatoric stress distribution inherited from the previous tectonic history. However, our experiments show that generally the response is similar to the generic one presented here, which avoids the complications of previous tectonic history and thermal transients by zeroing the initial stress distribution and equilibrating the thermal structure prior to compression.

The vertical movements of the basement are tracked at points A, B, and C (Fig. 4). During the compressional buildup phase of initially elastic strains, the applied force rises quickly, and a low-amplitude, downward, elastic flexure of the basin develops. As the applied force levels out at ~4×10^{12} N/m parallel to the strike, deviatoric stresses reach the yield surface in the central zone of crustal weakness. Crust and sediment start to thicken here and produce incipient topography that is compensated in regional isostasy by downward flexure of the relatively strong upper mantle. This process further deepens the flexural basin, except in the inversion zone where it shallows owing to topography produced in the ongoing compression. With further compression, point A rises above its initial position to signal the existence of a central inversion zone. The flexural basin now has developed into two characteristic (symmetrical) marginal troughs, which are deepest near the inversion zone and shallow away from it (Fig. 4, model time 10 m.y.).

Upon cessation of straining at 10 m.y., deviatoric stresses in the lower crust and subcrustal lithosphere relax viscously, in effect decreasing the flexural rigidity of the lithosphere. This effect can be observed in the time evolution of the applied force, which decays exponentially after cessation of straining. The stress relaxation and a slight heating of the compressionally thickened lithosphere leads to postcompressional, buoyancy-driven inversion and erosion of the central inversion zone and the marginal basins. The amplitude of the secondary inversion vanishes toward the borders of the marginal troughs and depends on the general thermal regime and the relieving of compressional stresses.

At model time 50 m.y. (Fig. 4), approaching the final steady-state situation, the synrift sedimentary rocks are exposed in the central inversion zone symmetrically flanked by the postrift sequence. The marginal troughs still exist, but mainly because compaction of the pre- and synrift sedimentary rocks have created space. More than 1.5 km of rock has been eroded from the central inversion zone.

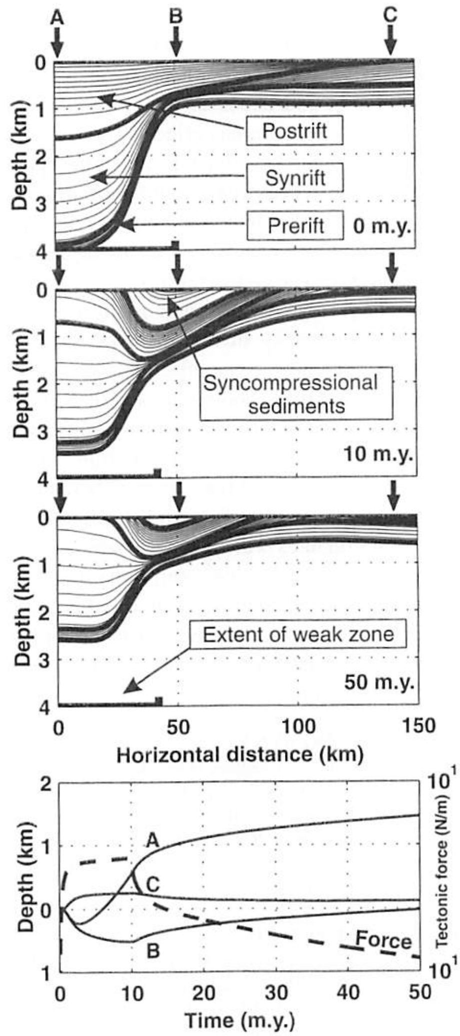

Figure 4. Stages in evolution of inversion structure. At model time 0 m.y.: Initial rift basin with pre-, syn-, and postrift sediments. Compression at constant rate lasts from 0 to 10 m.y. At model time 10 m.y.: Structure at end of compression. Marginal troughs are well developed and ~1 km of sediment has been removed from central inversion zone. At model time 50 m.y.: Approaching final steady-state situation. Marginal troughs and central inversion zone have been deeply eroded during postcompressional secondary inversion. Bottom: Time evolution of vertical deflection in points A, B, and C, and force present at model right boundary.

DISCUSSION AND CONCLUSIONS

In this model, the layered rheologic structure of the lithosphere is central to the formation and evolution of inversion zones and accompanying marginal troughs. The introduction of a weak zone defines where deformation takes place in extension or compression. Crustal shortening in the weak zone must be matched by an equal amount of mantle shortening. The mantle also shortens where it is relatively weak, which is not at the center of the rift but beneath the flanking basement highs, where the Moho temperature is at its maximum. The model therefore resolves the basin inversion enigma by connecting laterally offset zones of mantle and crustal shortening by a zone of shortening and shearing in the lower crust (Fig. 5).

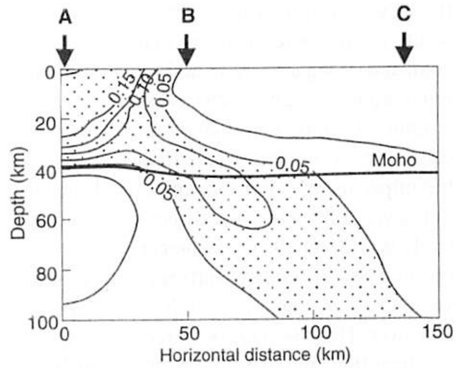

Figure 5. Strain contours at end of compression. Lithospheric deformation has occurred mainly in crustal weak zone and in upper mantle beneath flanking basement highs, where Moho temperature is highest.

The location and extent of this zone have a profound influence on the evolution of the marginal troughs, which occurs in a competition between two opposing effects. The load-induced flexure of the upper mantle deepens the trough, while any crustal shortening and thickening taking place outside the central inversion zone shallows the trough, mainly because of crustal thickening, but also because dispersing of the load on the upper mantle results in a smaller amplitude and a wider flexure of the mantle bulge.

If the weak zone comprises only the upper crust, the lower crust will shorten and thicken also under the marginal troughs and counteract their formation. Inclusion of the lower crust in the crustal weak zone is an efficient way to ensure that the lower crust thickens as little as possible under the marginal troughs. The fact that inversion zones tend to have marginal troughs perhaps implies that the majority of crustal shortening occurs in the weak zone and that the lower crust also must be part of the weak zone.

Many inversion zones show asymmetry in the marginal troughs. In some cases this may be caused by asymmetry in the lower-crustal shear zones connecting the zones of shortening in the crust and upper mantle. For example, the Sorgenfrei-Tornquist zone in the eastern North Sea area shows extreme asymmetry in the chalk depocenter (Figs. 1 and 2A). This asymmetry can be explained (Hansen et al., 2000) by a lower-crustal shear zone connecting the crustal shortening in the inversion zone to upper mantle shortening to the north under the Fennoscandian Shield. Crustal thickening in this shear zone eliminates the northern marginal trough. Contrary to this, there is no lower crustal shear zone reaching south because here the upper mantle is shallow, cold, and strong and does not shorten. The southern marginal trough therefore develops well.

Other asymmetrical marginal troughs can be caused by asymmetry in the flexural rigidity of the upper mantle. For example, the Tornquist-Teisseyre zone (Figs. 1, 2B) has wide and relatively shallow chalk depocenters to the northeast on the East European platform. To the southwest, the depocenters generally are deeper and narrower. This observation agrees well with the low heat flow and therefore relatively high flexural rigidity of the East European platform as compared to the laterally heterogeneous lithosphere of the younger and mobile Europe to the west. The larger thickness of the crust of the East European platform apparently is not sufficient to ensure a weak upper mantle. This effect may be entirely thermal or may exist because the upper mantle under the stable platform is different from (perhaps drier than) the upper mantle under western Europe.

If the weak zone reaches into the upper mantle, the amount of crustal thickening outside the weak zone is minimal. Asymmetric shearing of the lower crust is then not a likely candidate to produce asymmetric marginal troughs. The most likely remaining candidate is asymmetrical upper mantle (and crustal) flexural rigidity.

As discussed by Ziegler (1987), the Oslo Horn, and Glückstatt provide examples of grabens that show no sign of inversion. The fact that they formed indicates that they are zones of structural weakness. That they did not invert in the Late Cretaceous and Paleocene means that they were relatively stronger than the graben systems that did invert. We suggest that the reason may lie in the fact that they have not been the sites of any strike-slip movements during or after their formation (Ramberg, 1972; Kockel, 1995; Clausen and Korstgård, 1994). In contrast to this history, all of the inverted grabens of the Alpine foreland were created in extension with a strike-slip component (Mogensen, 1995; Pozaryski and Brochwicz-Lewinski, 1978; Vejbæk and Andersen, 1987; Lake and Karner, 1987; van Hoorn, 1987; van Wijhe, 1987; Betz et al., 1987). Strike-slip displacements are very efficient in creating many erratically oriented fault planes, thus ensuring that there are always some in a favorable position to become reactivated in transpression.

This model relates the process of inversion to zones of structural weakness, which produce localized shortening and thickening of the crust and sediments in compression. The topography of the inversion zone is regionally compensated by downward flexure of the relatively strong upper mantle, an effect that results in the formation of marginal basins synchronous with inversion and erosion in the central zone. The topography of the system can be understood as the superposition of two opposite topographic effects: The smooth, long-wavelength, downward flexure of the upper mantle caused by loading, and the rugged short-wavelength topography related to shortening and thickening of sediments and crust in the central inversion zone. The system relaxes after cessation of the compressional forces and inverts a second time, now regionally, involving both the central inversion zone and the marginal troughs.

Because localized shortening and thickening are fundamental to the development of inversion zones, they cannot be modeled by thin elastic plate models, which do not consider shortening. Further rheologic modeling of the evolution of inversion structures is likely to yield new insights into lithospheric rheology and dynamics and may, for example, shed light on the rheologic conditions under which the crust and upper mantle can preserve zones of structural weakness over geologic time.

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