Thermal and tectonic evolution of the southern Alps (northern Italy) rifting: Coupled organic matter maturity analysis and thermokinematic modeling

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ABSTRACT

The southern Alps were characterized by strong variations, both in space and time, of heat flow during Mesozoic rifting. The regional thermal history was reconstructed using organic matter (OM) maturity data from outcropping sediments. One-dimensional (1-D) thermal modeling performed on selected successions suggests that OM maturity was mainly controlled by high geothermal gradients (heat flow peaks of 85 to 105 mW/m² in the Middle Jurassic) and differential burial during Norian–Early Jurassic extensional phases. The results of 1-D modeling show an eastward increase of heat flow peak values.

These results were compared with those obtained with two-dimensional (2-D) thermokinematic models. The models show a time shift (ca. 10 Ma) in the heat-flow peak (Aalenian-Bajocian for 2-D and Bajocian for 1-D modeling). However, the Bajocian age was a priori imposed on 1-D models. Available geochemical data could be fitted assuming Aalenian-Bajocian peak ages. Consequently, this misfit is not alarming. The eastward increase in heat-flow peak values is tentatively explained with an eastward increase of radiogenic heat production in the crust instead of with differential stretching.

The comparison of paleothermal data and numerical modeling was done to gain knowledge on the potentials and limitations of numerical modeling in frontier areas. Although some
differences do exist in the results of geochemical and thermokinematic models, we can conclude that if a reasonable knowledge of the thermal parameters of both covers and basement is available, thermokinematic modeling can provide useful first-order estimates in frontier areas of heat flow and temperature evolution through time.

INTRODUCTION

The thermal history of sediments is commonly one of the main unknown variables in geochemical modeling aiming to evaluate the timing of hydrocarbon generation and expulsion. Assessment of the heat-flow history is fundamental in reconstruction of the chronological link between the age of structural development and generation and expulsion of hydrocarbons. In areas characterized by a good coverage of organic matter (OM) thermal maturity data, measured in boreholes or in outcrops, the geochemical models are well constrained. In such models, vitrinite reflectance (Ro) data are normally used for calibration. However, in many areas, paleothermal estimates, when available, are restricted to wells and lack lateral continuity. In such unconstrained or underconstrained areas, the prediction of formation temperatures can only be obtained by thermokinematic numerical models. In this way, estimating the heat-flow and temperature evolution through time in frontier areas characterized by a lack of wells and subsurface data becomes possible. In this latter case, uncertainties in kinematic and thermal parameters can heavily hinder the use of modeling results.

In this work, we address the thermal evolution of the Permian–Aptian succession of the southern Alps (northern Italy, Figures 1, 2) by means of field and laboratory thermal analyses and thermomechanical numerical modeling. The thermal record of the Mesozoic extensional tectonics obtained from analyses of OM maturity and 1-dimensional (1-D) geochemical modeling (e.g., Fantoni and Scotti, 2003; Scotti, 2005; Scotti and Fantoni, in press) is compared with the results obtained with a two-dimensional (2-D) thermokinematic model (see Grigo and Schmalholz, 2003, for a description of the adopted code). Both 1-D geochemical models and 2-D thermokinematic models have pros and cons. Two-dimensional thermokinematic models provide a solution that is regionally self-consistent and consider lateral heat flux, obviously neglected in 1-D models. However, 2-D models require detailed knowledge of stratigraphy and fault distribution along regional transects, and this kind of information is not always available in frontier areas. Moreover,
Figure 1. (a) Regional tectonic sketch of southern central Europe. The dashed rectangle shows the location of the map of panel b. The black areas show magmatic bodies. (b) Geological sketch of northern Italy. Simplified from Bigi et al. (1990). The black thick line is the trace of the profile of Figure 2.
because of the regional scale of such 2-D models and because of computational costs that limit the number of distinguishable formations, the resolution of the resulting thermal evolution is not sufficient for exploration and production geoscientists. On the contrary, 1-D models allow us to better reproduce available thermal data for single locations because the effects of single parameters are more easily predictable and the resolution is normally higher because computational costs are much lower, even considering detailed stratigraphies. However, 1-D modeling results obtained for single boreholes or locations are not necessarily regionally consistent. A common problem for 1-D and 2-D models is that, even if thermal (e.g., borehole temperature and heat flow) and geochemical (e.g., Ro) data are available for comparison with modeling predictions, the solution to the thermal problem may be nonunique. Such nonuniqueness is usually because thermal and, eventually, kinematic parameters are generally underconstrained. These problems, well known and of fundamental importance in oil geology, are tackled in this work. In particular, we test the effects of the main thermal and kinematic parameters on the thermal evolution predicted by 2-D models. This sensitivity analysis can be extrapolated to other cases and is expected to be valuable both for thermal modelers and for users of thermal modeling results. Besides addressing specific aspects of the thermal evolution of the southern Alps rifting at a regional scale, our comparison of paleothermal data and numerical modeling provides some useful suggestions on the potentials and limitations of numerical modeling in frontier areas.

GEOLOGICAL FRAMEWORK, STRATIGRAPHIC AND TECTONIC EVOLUTION

The southern Alps (Figure 1) are a low- to non-metamorphic south-vergent fold and thrust belt separated from the north-vergent Alpine, mainly metamorphic, orogen by the Insubric line, a dextral transpressive fault active since the Oligocene (Schmid et al., 1989). The geological units that form the southern Alps (Figures 2, 3a) were part of the Mesozoic passive continental margin of the Adriatic plate during the opening of the western Tethys (or Ligurian–Piedmont oceanic basin).

Petrological and structural data collected in the basement of the central southern Alps allow us to constrain the thermomechanical history and to link it to the geodynamic evolution of the area (e.g., Bertotti et al., 1993b; Siletto et al., 1993). The oldest phase of deformation recorded in the Orobie Alps basement is associated with amphibolite facies metamorphism and occurred during the Variscan orogeny (ca. 330 Ma; e.g., Siletto et al., 1993). This tectonometamorphic event was followed by green-schist facies retrogression. This pressure-temperature-time evolution is associated with the late Paleozoic Variscan subduction collision (e.g., Di Paola and Spalla, 2000).

Pennsylvanian–Middle Triassic

At the end of the Variscan orogeny (Pennsylvanian–Late Permian), widespread calc-alkaline magmatism and extensional (probably transtensional) tectonics affected the area of the southern Alps, as well as other large areas of Europe. Carboniferous–Lower Permian successions, locally beginning with conglomerates, are characterized by continental and lacustrine sediments and volcanics. The thickness of these successions is tectonically controlled. In the Late Permian, a predominantly clastic succession was deposited unconformably over the metamorphic basement and the Lower Permian volcanic, volcanioclastic, and sedimentary rocks. The analysis of metamorphism of lower and middle crustal rocks of the Ivrea zone and of the central southern Alps suggests the occurrence of a positive thermal anomaly coupled with extensional tectonics in the Permian–Early Triassic (Bertotti et al., 1993b).

The origin of the Late Carboniferous–Early Permian sedimentary troughs is still a matter of debate. They may be the result of late Variscan dextral wrenching, which affected large parts of southern Europe (e.g., Arthaud and Matte, 1977), or of postcollisional collapse of the Variscan belt (Malaveille et al., 1990). Other authors interpret these structures and the associated volcanic deposits as the initial stages of the Tethyan rifting (e.g., Winterer and Bosellini, 1981; Siletto et al., 1993).
The Early Triassic was generally characterized by transgression and deposition of sabka and shallow-marine deposits, with increasing thickness from west to east. In the Middle Triassic and early Carnian, differential motions led to the development of a complex pattern of carbonate platforms and intervening basins, with widespread volcanism in the Dolomites and below the present-day Po Plain. The generalized subsidence and the related rise of sea level, the lateral variations in thickness, and the deepening upward of the facies all support the presence of active extensional tectonics in the southern Alps during the Triassic. In addition, several synsedimentary normal faults of Early, Middle, and Late Triassic age occur in the Dolomites (e.g., Doglioni and Carminati, 2008).

The southern Alps Permian–Triassic stretching has also been interpreted as a relic of a back-arc basin in the hanging wall of a west-directed subduction zone (Doglioni, 1995).

Late Triassic–Early Cretaceous

In the Carnian, a major sea level fall coupled with the input of siliciclastic and volcaniclastic material caused the extinction of most of the platforms and the onset of sedimentation in a fluvio-deltaic system. This stage was followed by a new marine transgression and deposition of the late Carnian continental-evaporitic platform. In the Norian, a carbonate platform environment existed over the entire southern Alps. Sedimentation was accompanied by diffuse normal faulting, recognizable mainly by thickness changes and only locally by facies changes (Bertotti et al., 1993a). This tectonic phase was interpreted as the beginning of the rifting that led to the opening of the Ligurian-Piedmont ocean (Bertotti et al., 1993a) or as an earlier unrelated event (Bernoulli et al., 1990; Cozzi, 2000; Berra et al., 2008).

In the Lombardy Basin, at the beginning of the Rhaetian, an increased water depth coupled with a climatic change caused the extinction of the Norian carbonate platform and the deposition of clay-rich sediments. Carbonate platform sedimentation resumed only in the middle Rhaetian. In the eastern parts of the southern Alps, the Rhaetian was mainly characterized by shallow-water carbonate deposition. According to Bertotti et al. (1993a) and Manatschal and Bernoulli (1999), extensional tectonics were active also in the Rhaetian in the Lombardy Basin area, but the fault scarps had no morphologic expression.
Figure 3. Results of the 1-D geochemical modeling. (a) Stratigraphic column for the Iseo W area. The thick vertical bar indicates the vertical extent of the section analyzed in panel d. Czo = Calcare di Zorzino Formation. (b) Heat-flow distribution at basement reconstructed for four locations (see Figure 2 for their location) where sufficient thermal constraints were available. Notice the rather continuous increase of the peak value of heat flow from west to east. (c) Burial (left) and thermal (right) history for the Iseo W location associated with the thermal maturity model of panel d. The two burial curves refer to the top and bottom of naphthogenic Upper Triassic units (Calcare di Zorzino Formation). The temperature curves refer to the same layers and are calculated assuming alternatively the variable heat flow of panel b (solid curves) or a uniform heat flow through time (dashed curves). (d) The curves show vitrinite reflectance predicted by the 1-D thermal model for the Iseo W location, assuming the heat-flow distribution shown in panel b (solid curve) and a uniform heat flow through time (dashed curve). Notice the excellent fit between predicted and measured values. See Scotti (2005) for further details on the modeling assumptions.
because of the high sedimentation rates (similar geometries were described elsewhere by Doglioni et al., 1998).

A major extensional phase occurred in the early Early Jurassic, with an east-west-oriented stretching (north–south extensional faults). Several parts of the late Rhaetian carbonate platforms were drowned during the early–middle Early Jurassic, and hemipelagic and pelagic sequences were deposited in depressed basins. On structural highs, shallow-water carbonate sedimentation continued. The following alternation structural lows (basins) and relative highs developed from west to east (e.g., Bernoulli, 1964; Bosellini, 1973; Bertotti et al., 1993a): the Cusio–Biella–Canavese zones, the Lombardy Basin, the Trento high, the Belluno Basin, and the Friuli platform (Figure 2).

Early Jurassic faults developed with listric geometries, rooted at middle upper crustal depths (approximately 15 km [9 mi]) (e.g., the Lugano fault, to the west the Monte Generoso Basin, Figure 2) (Bertotti, 1991). The extensional tectonic activity culminated in the Hettangian–Sinemurian in the Lombardy Basin (representing the proximal area of the future passive margin) and ceased at the Pliensbachian–Toarcian boundary. In the Pliensbachian–Toarcian, stretching migrated more to the west to the distal parts of the future passive margin (Biella and Canavese zones; Berra et al., 2008). In these areas, normal faulting dissected the entire crust (Hodges and Fountain, 1984; Zingg et al., 1990). Extensional tectonics was associated with the exhumation of the lower crust in the Canavese and in the Ivrea zone (Hunziker, 1974; Ferrando et al., 2004).

Early Jurassic stretching has been related to the sinistral motion between the African and European plates (e.g., Channel et al., 1979). The rifting stage came to an end in the late Early Jurassic, when normal faulting in the southern Alps ceased and the oceanic crust of the Ligurian-Piedmont Ocean began to form (the oldest documented Penninic oceanic crust is Bajocian in age; Bill et al., 2001).

Post breakup subsidence was generally related to cooling processes, although the occurrence of syndepositional normal faults provides evidence of extensional activity also in the Late Jurassic–Early Cretaceous (Doglioni, 1992a, b). Since the end of the Early Jurassic, pelagic sediments, calcarenites, and mud turbidites have been deposited, with the exception of the western parts of the southern Alps (Canavese zone), where locally breccias, conglomerates, and sands are intercalated with basinal sediments, testifying to the occurrence of tectonic activity. In the Trento Plateau, sedimentation has been mainly condensed pelagic since the drowning of the carbonate platform in the Aalenian.

**Late Cretaceous–Pleistocene**

In the Late Cretaceous, the drifting of the Ligurian-Piedmont Ocean was interrupted by the onset of a subduction zone, which brought, in the Eocene, the closure of the ocean and the Alpine continental collision (e.g., Doglioni and Bosellini, 1987; Polino et al., 1990; Schmid et al., 1996). In the southern Alps, the subduction and collision stages were accompanied by the deposition of siliciclastic sediments with laterally variable thickness (e.g., Bersezio and Fornaciari, 1988; Bernoulli and Winkler, 1990).

Compressional deformations continued until the Miocene in the central-western sector of the southern Alps belt and in its Po Plain foreland (Fantoni et al., 2004) and until the Pliocene in the eastern sector (Castellarin and Cantelli, 2000). The related foredeep sediments, consisting of thick clastic sections, are preserved at the southern margin of the belt and in the Po Plain subsurface (Bersezio et al., 1993, for the western sector; Massari et al., 1986; Fantoni et al., 2002, for the eastern sector). Local compressional seismicity occurs today.

**Modeled Cross Section**

Figure 2 shows the reconstructed geometry of the southern Alps passive margin of the Adriatic plate in the Aptian. The sedimentary succession is divided into four major intervals: Permian–Carnian (299–216 Ma), Carnian–Rhaetian (216–200 Ma), Hettangian–Bajocian (200–168 Ma), and Bathonian–Aptian (168–118 Ma). Figure 2 is approximately referred to the trace shown in Figure 1, but it has been built with data coming from different sectors of the southern Alps belt. In fact, Permian–Triassic sediments crop out in the most internal sectors of the southern Alps.
the chain, whereas Upper Triassic–Lower Cretaceous sediments are observed in the central and southern sectors.

Only the major normal faults occurring in the area are shown in Figure 2. They were identified either by direct field recognition or by lateral thickness and facies changes. These faults are not included as input features in the models. In this way, the software used for thermokinematic modeling is left free to identify the loci of potential syndepositional faults on the basis of thickness changes.

Since the Cenomanian, siliciclastic turbidite systems, associated with the onset of shortening, have been deposited. These sediments are not included in the cross section of Figure 2 and are neglected in the calculations to separate the riftting and postrifting thermal evolution from the following orogenic stages. These deposits were neglected also because of the incompleteness of information on their thickness because in many locations, the sedimentary succession was later uplifted and strongly eroded. Their inclusion would have added much subjectivity in the modeled cross section. The thermal effects of the post-Aptian sedimentation will be, however, broadly covered in the following discussion.

GEOCHEMICAL MODELING

Various thermal histories have been described in the last few years for the southern Alps (Fantoni and Scotti, 2003; Scotti, 2005; Zattin et al., 2006; Scotti and Fantoni, in press). They were reconstructed using OM maturity values obtained using samples collected from Permian to Cretaceous sedimentary units cropping out along the whole chain (Figure 2). Locations with data clearly affected by anomalous thermal perturbations (e.g., deposition of thick foredeep clastics subsequently eroded; magmatic intrusions) were excluded to simplify analysis and interpretation. Because our purpose is the study of the kinematic and thermal evolution of the southern Alps during rift and postrift stages, lateral thermal overprints are unwanted because they may drive to misinterpretations. Particular care was devoted to exclude data obtained from successions doubled by thrusting. Thrusting could have induced a sudden burial and hence heating of the succession. Also, doubled successions exhumed by later erosion were excluded.

Almost all OM maturity data are from intervals with fair to good total organic carbon (TOC), ranging in most samples between 0.4 and 2%. In some localities (in particular near Iseo W), the presence of several high TOC levels provides a vertical assemblage of data along a thick sedimentary succession, providing good vertical definition of the maturity gradient. Vitrinite reflectance values from the western-central sector of the southern Alps (from the Biella zone to the Trento Plateau) are described by Calabrò et al. (2003), Fantoni and Scotti (2003), Bersezio et al. (2005), and Scotti (2005).

The lowest OM maturity values occur on structural highs characterized by thin overlying sedimentation. On the Gozzano and Arzo highs, for example, the maximum temperature ($T_{\text{max}}$) from Rock-Eval pyrolysis for Middle Triassic units is less than 425°C (immature kerogen stage), suggesting maximum temperatures lower than 60 to 70°C (using a heating rate of 3°C/m.y.). Also, samples from the Trento Plateau show low OM maturity ($R_o \sim 0.5\%$ and $435°C T_{\text{max}}$) for Upper Triassic units. An increase in OM maturity can principally be attributed to a thicker synrift sedimentary succession. The highest maturity values correspond to the depocenters of synrift successions, such as the Upper Triassic units of the Monte Generoso Basin and Iseo Basin (Iseo W, location in Figure 2) successions. Very high OM maturity was reached by Norian units ($R_o \sim 3.5\%$ at the base of the Calcare di Zorzino Formation, Figure 3a) in the Iseo W depocenter.

Vitrinite reflectance data were used to calibrate 1-D numerical models (proprietary Eni S.p.A.) simulating the degree of maturity reached by kerogen (for further details, please refer to Scotti, 2005; Scotti and Fantoni, in press). The model input data are (1) thickness, (2) lithology (compaction curve and thermal conductivity of the rock matrix), (3) age of sedimentation and paleobathymetry for each stratigraphic event (all of these data define the burial history), (4) heat-flow values through time, and (5) paleolatitude, which is used to convert paleo-surface temperature to sea-bottom temperature.
according to paleobathymetry. Finally, the input for the kinetic model has to be defined (e.g., using the method of Sweeney and Burnham, 1990). The distribution of heat flow through time is the main unknown variable. In the literature, several hypotheses have been proposed for the timing and magnitude of the heat-flow peak: 75 mW/m² in the early Liassic according to Novelli et al. (1987); about 90 mW/m² at Anisian–Ladinian and Carnian–Norian ages according to Greber et al. (1997) and Bertotti et al. (1993a), respectively.

In our analysis, best-fit solutions between calculated and measured OM maturity for some of the burial history models (i.e., Val Breggia; Iseo W; Non Valley, Agordo, and other simulation points in the eastern southern Alps) were obtained by assuming increased heat flow throughout the rifting stage, with a maximum during the Bajocian (Figure 3b). Resulting heat-flow values are quite high and relatively consistent throughout the southern Alps (85 to 105 mW/m²; Fantoni and Scotti, 2003). We stress here that the results of modeling show a rather continuous increase, from west to east, of the heat-flow values at the Jurassic peak. Such variations could be caused not only by differential crustal thinning, but also other causes such as differential radiogenic heat production within the crust or underplating phenomena. Heat flow progressively decreased after the Bajocian–Bathonian to values similar to the present day by the end of the Cretaceous. These reconstructions are consistent with the known tectonic evolution of the Mesozoic extension in the southern Alps, characterized by a rifting stage until the Early Jurassic followed by a drifting stage and thermal subsidence from the Middle Jurassic.

An update of the previous model was made for the most significant locality (Iseo W; Scotti, 2005; Scotti and Fantoni, in press), which is constrained by a wealth of OM maturity data to better define the heat-flow history. Figure 3d shows the maturity data for the upper part of the succession of this area and the results of the thermal model. To reproduce the relatively high OM maturity of the topmost 1.5 km (0.9 mi) of the succession, an overburden caused by the eroded Cretaceous–Paleocene succession had to be included (Scotti, 2005). Because of the erosion, no direct knowledge of this succession is available. A 2200-m (7218-ft) thickness of the Cretaceous–Paleocene succession has been estimated in the depocenter of the foredeep basin of the Eo-alpine chain (Bersezio et al., 1993). Even by assigning this upper bound thickness (2200 m [7218 ft]) to the eroded Cretaceous–Paleocene succession, the temperature increase resulting from 1-D modeling was insufficient to obtain a good match between modeled and measured maturity data. Scotti (2005) and Scotti and Fantoni (in press) proposed that this misfit could be solved by introducing a heat-flow decrease in the Cretaceous–early Tertiary less pronounced than previously assessed (Figure 3b, c).

**THERMOKINEMATIC MODELING**

**Thermokinematic Software**

The calculations were performed with a code (Grigo and Schmalholz, 2003), which numerically simulates sedimentary basin formation and is coupled with an automatic inversion algorithm. For a detailed description of the algorithm, please refer to Rüpke et al. (2008).

The software includes two different modules. The first is a numerical forward module that simulates extensional basin development and related sedimentation. It is based on pure shear kinematics (McKenzie, 1978; Kooi et al., 1992; Liu and Ranalli, 1998) and it allows for depth-dependent stretching (Royden and Keen, 1980; Hellinger, 1983) and multiple rifting events of finite duration (Jarvis and McKenzie, 1980; Reemst and Cloetingh, 2000).

The second is an automatic inversion algorithm that, through an iterative process, searches for the optimal set of stretching factors of the crust and the mantle (β and δ) and paleowater depths that minimize the misfit between observed and modeled stratigraphy. The 2-D inversion algorithms generally find the optimal set of thinning factors within 10 to 20 iterations for given initial conditions. In our case, the observed stratigraphy is shown in Figure 2, which is referenced to the Aptian. The optimal model must also adequately fit the geometry of the base of the Moho. This requires assumptions
for the crustal structure of the analyzed area at the Aptian, as discussed below.

The optimal set of parameters obtained by the iterative process is then used by the forward model to reconstruct the thermal history (heat flow and temperature). The thermal predictions (in particular heat flow and vitrinite reflectance) will be compared with available thermal data.

Main inputs are the timing of extensional phases, the initial (Permian) crustal and lithospheric thicknesses, and mechanical parameters such as depth of necking and effective elastic thickness. The adopted inputs will be discussed and justified in the following sections.

**Crustal Structure**

The southern Alps are covered by a network of reflection (vertical and wide angle) and refraction seismic profiles that allowed the construction of several isobath maps of the Moho discontinuity (e.g., Nicolich and Dal Piaz, 1990; Kissling, 1993; Scarascia and Cassinis, 1997; Waldhauser et al., 1998; Nicolich, 2001; Cassinis et al., 2003). Based on the same data set, several crustal transects have been constructed (e.g., EGT: Ye et al., 1995; NFP20: Schumacher, 1997; TRANSALP: Gebrande et al., 2002; Scrocca et al., 2003).

According to these maps, only little variations in present-day crustal thicknesses can be recognized along the considered profile. For instance, one of the most reliable Moho reconstructions in the study area (Figure 4) (Waldhauser et al., 1998) highlights a present-day slightly deeper Moho below the Trento Plateau (about 36 km [22 mi]) and a shallower Moho below both the Belluno (32 km [20 mi]) and Lombardy basins (30–32 km [19–20 mi]).

The restoration of the present-day Moho geometry to the Aptian geometry is accomplished by subtracting the effects of the tectonic thickening associated with the convergent stages of the Alpine cycle. For this purpose, we have defined in detail the
thickness of both upper and lower crust (UC and LC, respectively) at the intersections between published crustal transects and our profile. Crustal thickness ranges from about 30 km (about 20 km [12 mi] of UC and 10 km [6 mi] of LC) along the TRANSALP profile to 23 km (14 mi) (about 13 km [8 mi] of UC and 10 km [6 mi] of LC) along the EGT transect. Along these transects, the effects of the Alpine thickening were evaluated and extrapolated to the remaining points of our Aptian transect. Subtracting such thickening effects to the present-day crustal thickness, we reconstructed a speculative Moho depth for the Aptian. The Aptian Moho (Figure 2) is shallower than the present day and may well reflect the crustal thickness at the end of the rifting stage and prior to the onset of the Alpine thickening.

The prerifting (Permian) lithosphere geometry (thicknesses of the UC and LC and of the lithospheric mantle) must be imposed on the model as well. After several runs, in which we varied these input parameters, we assessed that the optimal initial thicknesses are as follows: upper crustal thickness, 25 km (15 mi); lower crustal thickness, 20 km (12 mi); lithospheric mantle thickness, 70 km (43 mi). These parameters allow the best fit between the crustal structure predicted by the model and the speculative Aptian Moho geometry shown in Figure 2. The thickened lithosphere, and the slight crustal thickening, assumed in the calculations is consistent with the fact that the Permian age was the final stage of the Variscan orogeny, which affected the southern Alps area.

**Timing of Extensional Deformation**

Scientists generally agree that main extensional phases in the southern Alps developed from the Late Permian to Middle Jurassic (e.g., Doglioni and Carminati, 2008). However, some evidence of extensional activity has been described also for the Late Jurassic–Early Cretaceous (Doglioni, 1992a, b), as also suggested by thickness variations in Figure 2.

Because the software models basin formation along a selected profile, basins generated by trans-tensional tectonics must be simulated as purely extensional basins. As a consequence, in our models, we have assumed as synriff all the units deposited between the Carnian and Aptian. Moreover, taking into consideration the uncertainties about the geodynamic significance of the Late Permian–Middle Triassic tectonic event, we have chosen to consider the Late Permian–Middle Triassic sedimentary units as synriff deposits as well. The model was therefore allowed to develop extensional tectonics throughout the entire period modeled.

In this study, we have adopted, as a time reference, the International Stratigraphic Chart (ISC) released by the International Commission on Stratigraphy (ISC, 2009). The main modeled horizons, and the corresponding absolute ages, are shown in Table 1.

### Rheology and Mechanical Parameters

To evaluate and assume mechanical parameters, such as the depth of necking or effective elastic thickness (EET) along the southern Alps cross section, rheological profiles were built. In the profiles (Figure 5), the yield stresses (in terms of differential stress $\sigma_1-\sigma_3$ needed to achieve deformation with either brittle or plastic mechanisms) are plotted versus depth (e.g., Ranalli and Murphy, 1987).

In agreement with previous rheological studies, we use Sibson’s law (Sibson, 1981) to calculate the brittle yield stress. This law, based on the Coulomb-Navier frictional criterion, predicts a linear increase of yield stress with depth:

$$\sigma_1 - \sigma_3 = \beta \rho g z (1 - \lambda)$$

where $\beta$ is a parameter depending on the fault type, $\rho$ is the density, $g$ is the gravity acceleration constant,

<table>
<thead>
<tr>
<th>Time Step No.</th>
<th>Stratigraphic Age</th>
<th>Absolute Age</th>
</tr>
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<tbody>
<tr>
<td>1</td>
<td>Aptian</td>
<td>118 Ma, end of calculations</td>
</tr>
<tr>
<td>2</td>
<td>Top Bajocian</td>
<td>168 Ma</td>
</tr>
<tr>
<td>3</td>
<td>Top Rhaetian</td>
<td>200 Ma</td>
</tr>
<tr>
<td>4</td>
<td>Carnian</td>
<td>216 Ma</td>
</tr>
<tr>
<td>5</td>
<td>Base Permian</td>
<td>299 Ma, beginning of calculations</td>
</tr>
</tbody>
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\[ s_1/C_0 \approx \varepsilon \left( \frac{\rho \Delta \sigma}{\sqrt{RT}} \right)^n \]

where \( s_1 \) is the depth, and \( \lambda \) is the pore-fluid pressure ratio (i.e., the ratio between fluid pressure and lithostatic load). In our calculations, the \( \beta \) parameter is assumed as 0.75 (typical of normal faults) and the average density \( \rho = 2800 \text{ kg m}^{-3} \). The fluid pressure is the least constrained parameter of the Sibson’s law. We assume that fluid pressure is close to hydrostatic (\( \lambda = 0.4 \)).

The applicability of the Coulomb-Navier criterion has been confirmed experimentally only down to midcrustal depths (Byerlee, 1978; Jaeger and Cook, 1979). Extrapolating it to greater depths could result in unrealistically high brittle strength. This observation is consistent with the experimental evidence that the coefficient of friction decreases with increasing confining pressure (Jaeger and Cook, 1979). Moreover, rock mechanic experiments show that high-pressure fracture mechanisms substitute, at depth, frictional failure (Shimada, 1993). These mechanisms are weakly dependent on pressure (Ranalli, 1997). Fracturing experiments on granites show that, at high pressures, the brittle strength becomes approximately constant (Shimada, 1993). Maximum differential stresses of about 200 MPa (29 ksi) can be sustained by granites (Shimada, 1993). Lithospheric mantle rocks are likely to support higher differential stresses (Fernandez and Ranalli, 1997; Ranalli, 1997). As a result, maximum strength of about 500 MPa (72.5 ksi) can be reasonably assumed for such mantle rocks.

The following power law creep relation (Carter and Tsenn, 1987) describes the ductile behavior.

\[ \sigma_1 - \sigma_3 = \left( \frac{\dot{\varepsilon}}{A} \right)^{1/n} \left( \frac{\rho \Delta \sigma}{\sqrt{RT}} \right)^n \] (2)

where \( \dot{\varepsilon} \) is the effective strain rate, \( A \) is the generalized viscosity coefficient, \( n \) is the stress power law.
exponent, $Q_c$ is the activation energy, $R$ is the ideal gas constant, and $T$ is the temperature.

Because the ductile rheology is mainly governed by temperature, we calculated lithospheric temperature profiles. The temperature distribution is calculated solving the unidimensional steady-state purely conductive heat transfer equation. Boundary conditions are the heat-flow densities at the surface and at the base of the lithosphere. The heat flow at the base of the lithosphere is calculated subtracting to the surface heat flow the heat flow produced by radioactive decay processes in the crust. The radiogenic heat production is assumed to occur only in basement crustal rocks and to decrease exponentially with depth.

The thermal diffusivity is taken as 2.5 W/mK for the sediments and 3.3 for remaining crustal and mantle rocks. The obtained heat flow at the base of the lithosphere is 30 mW/m², compatible with available regional estimates (Cèrmák et al., 1992). Table 2 shows the rheological parameters adopted in the calculations (Fernandez and Ranalli, 1997).

Figure 5a shows the rheological profiles obtained for surface heat flow of 60 mW/m², assuming that Sibson’s law is valid for the entire lithosphere. Figure 5b shows the rheological profiles obtained for the same surface heat flow assuming, however, that Sibson’s law is valid for the first 10 km (6 mi) of the crust and that, at greater depths, the brittle yield stress remains constant. In both cases, the lithosphere is composed of 25 km (15 mi) of UC, 20 km (12 mi) of LC, and 70 km (43 mi) of lithospheric mantle. Strain rate is assumed constant and equal to $10^{-14}$ s⁻¹. Such a configuration is consistent with that of the Adriatic lithosphere before the onset of rifting.

**Effective Elastic Thickness**

The EET of continental plates is characterized by great variability and is not easily correlatable to simple geodynamic parameters (Burov and Diament, 1995). The EET does not necessarily correspond to simple features such as the thickness of the brittle crust or brittle lithosphere (also named mechanical crust or mechanical lithosphere). Such a correspondence only occurs in cold (i.e., thermally old) continental plates where no ductile (i.e., weak) layers occur between the stiff continental brittle crust and continental lithosphere. However, the EET is a function of temperature, pressure, and mineralogical composition of the lithosphere. In other words, it is a function of the rheological stratification of the lithosphere. In thermally younger and rheologically stratified continental lithosphere, the EET ($T_e$) can be best calculated using the following relationship.

$$T_e = \sqrt[3]{h_1^2 + (h_2 - h_c)^3}$$

where $h_1$ is the thickness of the mechanically competent crust, $h_2$ is the thickness of the mechanically competent lithospheric mantle, and $h_c$ is the crustal thickness. Burov and Diament (1995) provided likely values of EETs of continental lithosphere as a function of thermal age (i.e., the age of the last tectonometamorphic event occurred in a certain region) and thickness of the lithosphere. Young lithosphere (with a thermal age of less than 100–200 m.y., such as the Adriatic lithosphere in the Permian–Triassic) is predicted to be characterized by EET around 15–20 km (9–12 mi).

From the rheological profiles of Figure 5 and using the numerical relations proposed by Burov and Diament (1995), we calculated the elastic thickness for the prerift and late-rift configurations. We obtained an EET of 24 km (14 mi) for the prerift and 9 km for the late-rift configurations. The value obtained for the prerift configuration is consistent with the predictions of Burov and Diament (1995) for young continental lithosphere. Moreover, this

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Upper Crust (Granite Wet)</th>
<th>Lower Crust (Diabase)</th>
<th>Lithospheric Mantle</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stress power law exponent ($n$)</td>
<td>3.2</td>
<td>3.2</td>
<td>4.5</td>
</tr>
<tr>
<td>Generalized viscosity coefficient $A$ (kJ/mol)</td>
<td>137</td>
<td>239</td>
<td>498</td>
</tr>
<tr>
<td>Activation energy $Q_c$ (MPa⁻ⁿs⁻¹)</td>
<td>$2.0 \times 10^{-4}$</td>
<td>$3.0 \times 10^{-2}$</td>
<td>$1.9 \times 10^5$</td>
</tr>
</tbody>
</table>

**Table 2. Rheological Parameters Used in the Calculations**
value is consistent with the 25-km (15-mi) value of the EET obtained by Bertotti et al. (1997) for the Adriatic lithosphere some 100 m.y. after breakup. A direct comparison between our and Bertotti et al.’s (1997) results is, however, hindered by the fact that the calculations are referred to very different times and geodynamic settings.

According to Burov and Diament (1995), the EET calculated with their equation can be diminished up to 50% by processes such as topographic load and horizontal forces. If the prerift-predicted EET (24 km [15 mi]) is accordingly corrected, the resulting elastic thickness is around 10–12 km (6–7 mi). It is concluded that, for the Permian–Triassic Adriatic lithosphere, reasonable values for the EET range between some 10 and 15 km (6 and 9 mi). The late-rifting stage is, however, characterized by a much lower elastic thickness (9-km [5.5-mi] EET, which, once corrected, could lower as much as to 4–5 km [2.4–3.1 mi]). In conclusion, a variable elastic thickness should be preferable to a constant elastic thickness. Once again, we performed a sensitivity analysis varying this parameter between 5 and 20 km (2.4 and 12 mi), as it will be discussed below. From such an analysis, we concluded that the optimal EET is approximately 15 km (9 mi).

Depth of Necking
The depth of necking is the level of no vertical motion in the absence of gravitational forces (Braun and Beaumont, 1989). Kooi et al. (1992) connected the level of necking with the zone of maximum lithospheric strength, located around the depth of the brittle to ductile transition. Further work (e.g., Spadini et al., 1995) showed that this relationship could be more complex when intracrustal or intralithospheric detachments occur.

Figure 5a and b both show that the maximum yield stresses in the crust are concentrated around a depth of 25 km (15 mi). Following the reasoning of Kooi et al. (1992), a depth of necking of 25 km (15 mi) should be assumed. Such rheological profiles were calculated assuming crustal and lithospheric thickness and surface heat flow consistent with the Permian prerifting situation. Rifting is expected to decrease the crustal and lithospheric thicknesses and to increase the surface heat flow. To evaluate the effects of rifting on the lithospheric strength, we calculated additional rheological profiles.

Figure 5c shows a rheological profile obtained for a surface heat flow of 80 mW/m², assuming that Sibson’s law is valid for the entire lithosphere. Figure 5d shows the profile obtained for the same surface heat flow assuming, however, that Sibson’s law is valid for the first 10 km (6 mi) of the crust. In both cases, the lithosphere is composed of 20 km (12 mi) of UC, 10 km (6 mi) of LC, and 50 km (31 mi) of lithospheric mantle. Strain rate is assumed constant and equal to $10^{-14}$ s⁻¹. Such a configuration simulates the strength of the Adriatic lithosphere after the onset of the rifting. From Figure 5c and d, we concluded that the maximum yield stress in the crust is concentrated around a depth of 8–10 km (4.9–6 mi), i.e., shallower than at the onset of rifting. This seems to indicate that the use of a depth of necking variable through time (following, for example, the base of the UC or of an isotherm) should be preferred to a depth of necking constant through time. This is consistent with the findings of Govers and Wortel (1999), who concluded that no one-to-one relationship exists between the necking depth parameter and the strength distribution in the lithosphere and that variations of the necking depth occur through time. Because such a variation is not possible with the adopted code, and given the uncertainties on this parameter, we performed a sensitivity analysis varying this parameter between 5 and 25 km (3.1 and 15 mi). From such an analysis, we concluded that the optimal necking depth is approximately 20 km (12 mi).

Rock Properties
The values that we adopted in our calculations for density, thermal conductivity, and radiogenic heat production rate are provided in Table 3. For sedimentary rocks, the parameters adopted in the calculations are commonly used in literature and are not constrained by measurements from rock samples from the southern Alps. The rock parameters for the UC and LC have been assumed considering measurements of petrophysical properties on rock
samples collected along the TRANSALP seismic reflection profile through the eastern southern Alps (Vosteen et al., 2003). Given the wide range of radiogenic heat production measured in southern Alps basement samples by Vosteen et al. (2003), this parameter has also been varied to evaluate the sensitivity of the model. No results are shown for sensitivity tests varying the rock conductivity. Our results are, however, compatible with those of Ter Voorde and Bertotti (1994), who showed that when the range of values used in literature is considered, the difference in temperature is less than 4%.

### Thermal Age Parameter

This parameter allows us to model an initial situation in which the lithosphere has not yet reached thermal equilibrium. This parameter is influenced by the age of the last tectonothermal event, which affected the study area. The software assumes that the entire lithosphere, at the end of a tectonothermal stage, has a constant temperature (1300°C) and then cools down for a time specified by the mentioned parameter. With the standard value of 2000 m.y., the lithosphere would have reached a steady-state condition at the beginning of the model. However, given geological evidence, one might want to include the effects of a previous rift or plume. In this case, the thermal age parameter should be set to a value that is less than the default.

In the southern Alps, the peak of the Hercynian thermal imprint related to the convergent stages is dated some 100–200 m.y. before the onset of the Alpine rifting stages that we are modeling. The time span since the rifting related to the Hercynian cycle is much larger. We performed a sensitivity analysis on this parameter and found that the influence of this parameter on the thermal setting of the alpine rifting cannot be neglected for thermal ages smaller than 20–30 m.y. For greater thermal ages, these effects are negligible. For this reason in our calculations, we used the default value for this parameter.

### Modeling Results

We performed modeling runs adopting the following criteria.

1. A first set of tests was performed to check the effects of the variation of mechanical parameters (depth of necking and EET) on the modeled stratigraphy and crustal structure.
2. A second set of tests (not shown) was conducted to evaluate the effects of the variation of thermal parameters (e.g., thermal rock properties and thermal age of the lithosphere).

Based on the outputs of these tests, it was possible to determine an optimal set of parameters. Finally, the heat flow and thermal evolution though time at the selected location obtained with the model built on the optimal set of mechanical and thermal parameters were cross checked against the thermal constraints (R_o and heat-flow evolution) provided by the analyses of the OM maturity.

### Effects of the Effective Elastic Thickness

Figure 6a shows the effect of the variation of the EET on the modeled crustal architecture and sedimentary basin shape and on the heat-flow distribution at the basement top in the Aptian (runs performed with a fixed necking depth of 20 km [12 mi]). Note first that all the models manage to reproduce the long-wavelength shape of the sedimentary basin (both top and bottom). However, observed short-wavelength peaks of the bottom of the sedimentary basin could not be fitted because only smoothly varying crustal and mantle stretching factors are allowed in the model. As the Moho geometry is concerned, very low EET values (e.g., 5 km [3.1 mi]) induce local compensation. As a consequence, the resulting Moho shape is rather wavy and mirrors the present-day basement

<table>
<thead>
<tr>
<th></th>
<th>Heat Production (W/m³)</th>
<th>Thermal Conductivity (W/m/K)</th>
<th>Density (kg/m³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbonates</td>
<td>1 × 10⁻⁶</td>
<td>2</td>
<td>2600</td>
</tr>
<tr>
<td>Upper crust</td>
<td>Between 2 × 10⁻⁶ and 6 × 10⁻⁶</td>
<td>3.5</td>
<td>2700</td>
</tr>
<tr>
<td>Lower crust</td>
<td>1 × 10⁻⁶</td>
<td>3</td>
<td>2900</td>
</tr>
</tbody>
</table>

*Table 3. Thermal Parameters Adopted for the Thermokinematic Modeling*
If the elastic thickness is increased to 10–20 km (6–12 mi), the Moho geometry becomes smoother. Following theoretical considerations, it has already been suggested that the elastic thickness for the Adriatic lithosphere is likely on the order of 15–20 km (9–12 mi). Remarkably, elastic thickness values of about 15 km (9 mi) produce a Moho geometry that resembles the presumed Aptian Moho. Obviously, the match of the modeled Moho to the presumed Aptian Moho is not proof that the model is correct. It just ensures its consistency with this set of data.

Figure 6b shows that the thermal effects of the variation of this parameter are rather minor. In general, a decrease of the elastic thickness is associated with an increase in the amplitude of oscillations of the curve of the heat-flow at the basement top.

**Effects of the Necking Depth**

Figure 7 shows the effects of the variation of the necking depth on the modeled crustal architecture and sedimentary basin shape and on the heat-flow distribution at the basement top in the Aptian. As shown also for the sensitivity analysis of the EET, all the models manage to reproduce the long-wavelength shape of the sedimentary basement (both top and bottom). The fit of the shape of the predicted Moho with the presumed Moho for the
Aptian is strongly variable. In particular, models with a shallow necking depth predict a very wavy Moho, at odds with the rather smooth shape of the Moho presumed for the Aptian. Only the models with deeper necking depths (in particular, with a necking depth of 20 km [12 mi]) can adequately reproduce the Aptian Moho shape. The optimal 20-km (12-mi) value for this parameter is consistent with the theoretical predictions discussed in a previous section. The thermal effects of the variation of this parameter (Figure 7b) are rather small. This suggests that the thermal part of the modeling is rather insensitive to this parameter.

Preferred Model

The geometric parameters used in the preferred model (listed in Table 4) were derived from theoretical

<table>
<thead>
<tr>
<th>Geometric Parameter</th>
<th>Thickness (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Initial upper crustal thickness</td>
<td>25</td>
</tr>
<tr>
<td>Initial lower crustal thickness</td>
<td>20</td>
</tr>
<tr>
<td>Initial lithospheric mantle thickness</td>
<td>70</td>
</tr>
<tr>
<td>Elastic thickness</td>
<td>15</td>
</tr>
<tr>
<td>Necking depth</td>
<td>20</td>
</tr>
</tbody>
</table>
Figure 8. Modeling results obtained for the preferred model assuming, for the upper crust, a radiogenic heat production of $4 \times 10^{-6}$ W/m$^3$. (a) Crustal geometry; (b, c) modeled and input stratigraphy, respectively; (d) distribution of heat flow at the top of the basement through time; (e, f) $\delta$ (crustal stretching) and $\beta$ (mantle stretching) factors, respectively. HFU = heat-flow unit.
considerations and from the sensitivity analysis described above.

As already shown in Figures 6 and 7, the crustal geometry predicted by the preferred model (Figure 8) mimics the reconstructed geometry of both the basin and Moho at the Aptian. Figure 8b and c show the modeled and input stratigraphy, respectively. Their long-wavelength shapes are quite similar, indicating that our preferred model simulates satisfactorily the southern Alps rifting. The short-wavelength oscillations of the sedimentary formations, induced by faulting, cannot, however, be simulated, as already discussed for the shape of the basin bottom.

Figure 8d shows the distribution of heat flow at the basement top through time along the modeled section. The interpretation of this graph needs to be coupled with that of Figure 8e and f that show crustal and mantle stretching factors, respectively. The initial heat flow is quite strong and constant throughout the basin, with values between 2.2 and 2.3 heat-flow unit (HFU). This is induced by the assumption of a 45-km-thick (28-mi-thick) crust characterized by quite high heat production and is consistent with the late Variscan emplacement of intrusive bodies. A major period of crustal stretching occurred between the Permian and Carnian (299–216 Ma), especially in the eastern and central parts of the basin. This stretching and the related fast sedimentation are associated with a decrease of heat flow (values between 2 and 2.2 HFU), especially in the easternmost sectors. This decrease is induced by sediment blanketing effects. The successive stage of stretching (Carnian–top Rhaetian, 216–200 Ma) is characterized by much lower crustal and mantle stretching factors and is characterized by generally lower heat-flow values (smaller than 2 in the same parts of the central and eastern sectors). Higher stretching factors characterize the following period (top Rhaetian–top Bajocian, 200–168 Ma), which is also characterized by an increase of heat flow (with values generally around 2.1–2.4 HFU). This increase of heat flow is induced by stretching-related mantle upwelling. The thermal effects of this process are shifted with respect to the time of stretching caused by the low thermal conductivity of rocks. The top-Bajocian–Aptian period (168–118 Ma) is characterized by progressively lower heat-flow values. Stretching in the Bajocian–Aptian stage is negligible, suggesting a phase of cooling and thermal subsidence in the basin.

Effects of the Radiogenic Heat Production
As shown in Figure 9, an increase or a decrease of assumed upper crustal heat production induces a corresponding increase or decrease of heat flow for the entire duration of rifting. Given the wide range of uncertainty on radiogenic heat production of southern Alps upper crustal rocks (Vosteen et al., 2003), we varied, in our preferred model, this parameter between $2 \times 10^{-6}$ and $6 \times 10^{-6}$ W/m$^3$. Figure 10 shows that such a variation produces dramatic changes (approximately 0.8 HFU, i.e., about 33 mW/m$^2$) in the Aptian heat flow at the top of the basement. Such changes are far beyond those considered reasonable in geochemical modeling performed to evaluate the timing of hydrocarbon generation and expulsion in rift-related basins. Evidently, when focusing on mostly unexplored frontier areas, a major effort should be dedicated to evaluate and constrain such thermal parameters. Moreover, a crosscheck with temperature profiles obtained from boreholes or with thermal data coming from OM maturity is crucial to model reliably the temperature evolution of a basin, as shown below.

COMPARISON BETWEEN 2-D THERMOKINEMATIC MODELING PREDICTIONS AND 1-D GEOCHEMICAL MODELING RESULTS

Figure 11 shows a comparison between the heat-flow evolution reconstructed by 1-D thermal modeling at the studied locations using OM maturity data and the heat-flow evolution predicted by our preferred thermokinematic model. Such a comparison shows a slight shift of the heat-flow maximum (Aalenian–Bajocian from 2-D modeling and Bajocian from 1-D thermal data). However, the Bajocian age for the thermal peak in the geochemical models was imposed a priori. The only available constraint for the peak age is that it has to follow
Figure 9. Distribution of the heat flow at the top of the basement through time along the modeled section for the preferred model with radiogenic heat production of the upper crustal rocks of $2 \times 10^{-6}$, $4 \times 10^{-6}$, and $6 \times 10^{-6}$ W/m$^3$. HFU = heat-flow unit.
the deep Early Jurassic burial. Available geochemical data could be fitted also assuming Aalenian–Bajocian peak ages. For this reason, we are not concerned about the misfit. Moreover, the change in the heat flow before and after the peak is more pronounced in the geochemical modeling than in the curves predicted by the thermokinematic modeling.

Apart from these differences, the thermokinematic model reasonably simulates the thermal evolution associated with rifting. The sensitivity analysis of the rocks’ thermal parameters (mainly the radiogenic heat production) showed that the heat-flow distribution, both in space and time, is strongly dependent on the assumed crustal heat production. Assuming a constant heat production over the entire model (as imposed by the software), a reasonable fit only in limited parts of the model is obtained. In fact, the higher values of heat flow are located in association with structural highs, whereas in the depocenters, our modeling reproduces the blanketing effect, yielding a strong decrease of heat flow at the base of the sedimentary succession. Progressively larger heat production is needed moving eastward to obtain a reasonable agreement between predicted heat flow and that calculated from available thermal data. In some simulations (not shown), we introduced also a process of underplating in the Ladinian in the eastern parts of the model to check if such a geodynamic process could be responsible for the higher heat flow in the eastern parts of the model. Such an underplating process was introduced to simulate the thermal event that drove the development of the Ladinian magmatism, rather widespread in the Dolomites, with minor evidence in the remaining parts of the southern Alps. These models did not indicate any significant difference in the Middle Jurassic heat flow with respect to the preferred model. Other causes, such as differential burial or crustal thinning between the western and eastern sectors, should be invoked to explain the heat-flow variations predicted by 1-D modeling. Both burial and crustal thinning are automatically calculated, in our 2-D models, on the basis of the input stratigraphy that, in principle (although the stratigraphy of the southern Alps is very well studied), could be locally unrealistic or wrong. The effects of uncertainties in the input stratigraphic architecture were evaluated with a set of models in which the bottom of the sedimentary cover was shifted uniformly downward or upward by 1000 m (3281 ft). Figure 12 shows the heat-flow histories predicted at a point located at a distance (along the $x$ axis) of 200 km (124 mi) (i.e., at the midpoint of the model). Results are that upward or downward shifts of 1000 m (3281 ft) of the bottom of the cover sequence determine an increase and a decrease of about 5 mW/m$^2$ of the heat-flow peak, respectively. Keeping fixed the crustal heat production at $2 \times 10^{-6}$ W/m$^3$, the heat-flow peak for the Val Breggia location is fitted, whereas a misfit of approximately 20 mW/m$^2$ occurs eastward (e.g., at Agordo). If such misfit is explained exclusively by differential stretching not reproduced by the models because of wrong input stratigraphy, then misplacements of several kilometers (3–5 km [1.8–3.1 mi] depending on the horizons in which the misfit occurred) of some of the stratigraphic horizons should have occurred. Given the high standard of the stratigraphic studies for the

![Image of Heat Flow Diagram](image-url)

**Figure 10.** Heat flow at the top of the basement at the end of modeling (118 Ma) predicted varying, in the preferred model configuration, the heat production between $2 \times 10^{-6}$, $4 \times 10^{-6}$, and $6 \times 10^{-6}$ W/m$^3$. HFU = heat-flow unit.
Figure 11. Comparison between the heat-flow evolution reconstructed at the studied locations using organic matter maturity data and the heat-flow evolution predicted by the preferred thermokinematic model, varying the radiogenic heat production of the upper crustal rocks between $2 \times 10^{-6}$ and $6 \times 10^{-6}$ W/m$^3$. 
southern Alps, this scenario is unrealistic. Because of the lack of any other reasonable explanation, the increase, from west to east, of heat-flow peaks reconstructed from OM maturity data is tentatively related to an eastward increase of the radiogenic heat production in the crust instead of to tectonic factors. This interpretation is in agreement with the strong variability in the nature of this parameter.

Figure 13 shows a comparison between the available measurements of $R_o$ (Fantoni and Scotti, 2003; Scotti, 2005; Scotti and Fantoni, in press) and the values predicted by our modeling (using the algorithm by Burnham and Sweeney, 1989; Sweeney and Burnham, 1990). The fit is reasonably good (given the regional scale of our study), especially for the model calculated assuming a crustal heat production of $4 \times 10^{-6}$ W/m$^3$. At places, model-predicted values are larger and at places smaller than available measurements. Clearly, such misfits cannot be resolved by increasing or decreasing model thermal parameters such as radiogenic heat production or varying the thermal parameters assumed for the sedimentary cover. The 2-D thermokinematic software does not allow the lateral variation of input parameters (with the exception of stratigraphy). As a consequence, the modification of these parameters to enhance the fit in one place would drive to a larger misfit in other regions.

Figure 14 shows a comparison between the $R_o$ measured by Scotti (2005) at the Iseo W location (Figures 2, 13) and the model predictions. The purpose of this figure is to show how different parameters locally affect the predicted $R_o$ instead of finding optimal input parameters for the entire model, as discussed before. In Figure 14a, the curves refer to models with the optimal set of kinematic and thermal parameters, with heat production of the upper crustal rocks ranging between $2 \times 10^{-6}$ and $6 \times 10^{-6}$ W/m$^3$ and with a sediment heat production of $1 \times 10^{-6}$ W/m$^3$. The predicted curves match quite well the trend of the measured $R_o$. However, the magnitude of the $R_o$ is generally underestimated, with smaller errors for the models characterized by the highest crustal heat-flow production. To investigate the effects of sediment heat production on the $R_o$ predictions, we ran two models (Figure 14b), increasing the sediments heat production to $2 \times 10^{-6}$ W/m$^3$. The misfit at deeper depths is reduced with respect to the models of Figure 14a, but the misfit at shallower depths is left unchanged.

In Figure 14c and d, the effects of the post-Aptian siliciclastic sediments on the $R_o$ predictions are shown. The thickness of post-Aptian sediments in the Iseo W area is not constrained because of erosion. As a consequence, we varied this parameter between 0 (as in all the previous models) and 2000 m (6562 ft) (an intermediate thickness of 1000 m [3281 ft] is a more reasonable estimate). Figure 14c and d show, at shallow depths, an improved match between predicted and measured values with respect to the results shown in Figure 14a and b, in particular for thicknesses of siliciclastic sediments of 1000 m (3281 ft). At depths deeper than 3 km (1.8 mi), the fit is excellent only for crustal heat production of $4 \times 10^{-6}$ W/m$^3$ and 1000 m (3281 ft) of post-Aptian sediments. The remaining curves, especially those obtained assuming 2000 m...
(6562 ft) of post-Aptian sediments, predict \( R_o \) values far in excess of available measurements. These observations allow us to conclude that the inclusion of the post-Aptian sediments is necessary to reproduce adequately the measured \( R_o \). The optimal values predicted for the ISEO W locality, in particular, the thickness of post-Aptian sediments, would not necessarily be adequate for the other areas because the thickness of such sediments was laterally highly variable. The comparison between measured and predicted \( R_o \) could be an important tool to reconstruct the thickness of eroded piles of sediments.

We, however, emphasized that the fit between model predictions and \( R_o \) data is reached assuming different overburden thicknesses and heating histories for geochemical and thermokinematic models. This observation suggests that no unique solution exists for the problem. The consequence is that the largest array of available observations should be used to constrain both kinds of models. This may not be possible in frontier areas. Therefore, in such areas, the results of both geochemical and thermokinematic models should be considered as first-order approximations of the real solution.
Figure 14. Along-depth comparison between the vitrinite reflectance measured by Fantoni and Scotti (2003) and Scotti (2005) at the Iseo W location (Figures 2, 13) and model predictions. (a) The various curves are referred to models with the optimal set of kinematic and thermal parameters, with heat production of the upper crustal rocks ranging between $2 \times 10^{-6}$ and $6 \times 10^{-6}$ W/m$^3$ and with sediment heat production of $1 \times 10^{-6}$ W/m$^3$. (b) The curves refer to models with the optimal set of kinematic and thermal parameters, with heat production of the upper crustal rocks ranging between $4 \times 10^{-6}$ and $5 \times 10^{-6}$ W/m$^3$ and with sediment heat production of $2 \times 10^{-6}$ W/m$^3$. (c) The curves show the results of models with the optimal set of kinematic and thermal parameters, with heat production of the upper crustal rocks of $4 \times 10^{-6}$ and with sediment heat production of $1 \times 10^{-6}$ W/m$^3$ varying the thickness of post-Aptian sediments between 0 and 2000 m (6562 ft). (d) The curves show the results of models with the optimal set of kinematic and thermal parameters, with heat production of the upper crustal rocks of $5 \times 10^{-6}$ and with sediment heat production of $1 \times 10^{-6}$ W/m$^3$ varying the thickness of post-Aptian sediments between 0 and 2000 m (6562 ft).
CONCLUSIONS

The OM maturity data and the results of 1-D thermal modeling were compared to the predictions of 2-D thermokinematic models based on the inversion of the Permian–Aptian stratigraphy of the southern Alps passive margin, restored at the Aptian by subtracting Cretaceous to present compressional structures (Figure 2). Irrespective of the differences in adopted conductivity values, allowing only qualitative comparison, 1-D thermal and 2-D thermokinematic models predict a slight shift of the heat flow maximum (Bajocian for 1-D and Aalenian-Bajocian for 2-D models) but are generally consistent. However, a sensitivity analysis on the rock thermal parameters adopted in 2-D models showed that the heat flow distribution, both in space and time, is strongly dependent on the assumed crustal heat production. A reasonable congruence between peak values predicted by 1-D and 2-D models is not obtained by imposing a unique value for heat production on the entire 2-D model. The increase, from west to east, of heat flow reconstructed from OM maturity data is tentatively related to an eastward increase in radiogenic heat production of the crust instead of tectonic factors (e.g., differential stretching).

Referring to the Iseo W case, the two models show different heat-flow peak values at the Middle Jurassic and different patterns of heat-flow decrease during the Late Jurassic and Cretaceous. This means, for the Triassic succession, maximum temperatures during the Middle Jurassic for the 1-D geochronological modeling and in the Cretaceous for the 2-D thermokinematic modeling. Given the influence of the thermal peak age on the timing of hydrocarbon generation, this misfit should be further investigated, including new calibration points. Again, Ro profiles are fitted by 1-D and 2-D models assuming different thicknesses of Cretaceous–Paleocene eroded sediments (2200 and ∼1000 m [7218 and ∼3281 ft], respectively). These results bear important consequences for the use of similar thermokinematic models in frontier areas. Without any reasonable knowledge of the thermal parameters of both sedimentary fill and basement and without a control of predicted temperatures and maturity data against borehole data, misfits between predicted and actual thermal evolution can occur. However, if at least the thermal parameters of both covers and basement are known, thermokinematic numerical modeling can provide a generally acceptable first-order estimate of the heat flow and temperature evolution through time.

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