3D structural and thermal modelling of Mesozoic petroleum systems in the Po Valley Basin, northern Italy

Claudio Turrini1*, Barbara Bosica2, Paul Ryan3, Peter Shiner-2, Olivier Lacombe4 & François Roure5,6

1 CTGeoConsulting, 78100, St Germain-en-Laye, France
2 Petroceltic International, Via Ennio Quirino Visconti 20, Roma 00193, Italy
3 Petroceltic International plc, 3 Grand Canal Plaza, Grand Canal Street Upper, Dublin 4, Ireland
4 Sorbonne Universités, UPMC Université Paris 06, CNRS, Institut des Sciences de la Terre de Paris (iSTeP), 4 place Jussieu, 75005 Paris, France
5 IFP-EN, 1 & 4 avenue de Bois-Préau 92852, Rueil-Malmaison, France
6 Tectonic Group, Faculty of Geosciences, Department of Earth Sciences, Utrecht University, Budapestaan 4, 3584 CD Utrecht, The Netherlands

*Correspondence: clturri@wanadoo.fr

Abstract: 1D and 3D basin modelling was performed to investigate the Mesozoic carbonate petroleum systems of the Po Valley Basin (northern Italy), through integration of a recent 3D structural model of the study area with the distribution of potential Triassic source rocks, rock properties and heat flow models.

Results from standard 1D maturity models show significant overprediction of the thermal maturity of deep Triassic carbonates in the western Po Valley, unless the effect of the substantial overpressure observed in these sequences is incorporated into the model. In order to further test this observation, two thermal scenarios were applied to the Po Valley 3D geo-volume: one based on the actual geological heat flow and a second model based on a reduced heat flow as a proxy for the delaying effect of overpressure on hydrocarbon maturation. The predictions of these two models were then compared with the observed hydrocarbon distribution in the western Po Valley.

Both thermal scenarios are broadly consistent with the observed hydrocarbon distribution at the scale of the basin but, in detail, the overpressure model provides a better match between the predicted charge available from the kitchen area’s post-critical moment and observed volumes of hydrocarbons initially in place within the traps, as well as with the observed and predicted hydrocarbon phases, as measured by the gas/oil ratio (GOR) of the fluids. Overpressure probably significantly delayed hydrocarbon maturation in the western domain of the basin, confirming results from previous studies.

Beyond regional implications, and despite its relative simplicity and inherent uncertainties, the adopted approach demonstrates the potential of a consistent 3D integration of the thermostructural history of sedimentary basins to constrain the geometry and structural evolution of hydrocarbon-bearing traps, as well as the generation and migration of hydrocarbons into these traps.

Received 7 March 2017; revised 24 August 2017; accepted 30 August 2017
hydrocarbon maturation and generation. Beyond regional implications, this study demonstrates the utility and applicability of an integrated 3D basin modelling approach to better constrain the geometry and structural evolution of hydrocarbon-bearing traps in sedimentary basins, as well as the generation and migration of hydrocarbons into these traps. Notably, the study confirms that the delaying effect of overpressure can be an important factor to be taken into account in predictions of hydrocarbon maturation and generation.

The Po Valley Basin

Regional geological setting

The geological architecture of the Po Valley Basin has been discussed in many recent papers covering the different structural–stratigraphic aspects of the region (e.g. Turrini et al. 2014, 2015, 2016 and references therein).

The Po Valley Basin is a complex basin system that developed as a nearly simultaneous pro/retro foreland-foredeep of the
diachronous and opposite-verging Northern Apennines and Southern Alps mountain belts. During Mesozoic and Cenozoic times, the Po Valley domain was affected by repeated extensional and compressional events (Fig. 1b). These tectonic events essentially relate to the long-lasting geodynamic effects produced by Tethyan rifting and drifting, and subsequent oceanic subduction and collision of the Adria and Eurasian plates (Dewey et al. 1973; Castellarin 2001; Carminati & Doglioni 2012; Pfiffner 2014 and references therein). Indeed, the present-day structural pattern is primarily a result of Mesozoic extension and Cenozoic compression (Pieri & Groppi 1981; Castellarin et al. 1985; Cassano et al. 1986; Bongiorni 1987; Fantoni et al. 2004; Ravaglia et al. 2006; Fantoni & Franciosi 2010; Turrini et al. 2014 and references therein). From Palaeogene to present times, the amplification and propagation of the Northern Apennines and Southern Alps belts controlled the differential flexure of the Po Valley–Adria lithosphere, the associated tilting and bulging of the foreland domain, the rapid sedimentation of thick foredeep-type deposits, and their successive involvement within the developing tectonic wedges (e.g. Carminati & Doglioni 2012 and references therein).

Mainly Miocene–Pleistocene thrusting is dominant across the shallow Tertiary sediments, whereas a large part of the basin substratum (Mesozoic and basement) shows evidence of the pre-compressional tectonic grain, with autochthonous highs and lows of extension-related origin partially reactivated by compression. Interference between the extension-related structures (approximately north-south trending) and the compression-related ones (generically west-east trending) is a primary characteristic within the basin (e.g. Turrini et al. 2016) that, given the earthquake distribution, is considered a more active tectonic province as one moves from west to east (Michetti et al. 2013; Vannoli et al. 2014; Turrini et al. 2015 and references therein).

The main stratigraphic units across the basin consist of Triassic platform carbonates and Jurassic–Cretaceous platform and basinal carbonates, overlain by Tertiary clastics (Fig. 1c) (Jadoul 1986; Catì et al. 1987; Jadoul et al. 1992; De Zanche et al. 2000; Ghielmi et al. 2012; Masetti et al. 2012; Pfiffner 2014). This sedimentary package appears to overlie some Permian sediments and their Hercynian metamorphic basement (Fig. 1c). The latter has been drilled by a few wells within the basin and locally crops out in the hinterland of the Southern Alps and the Northern Apennines (Cassano et al. 1986; Ponton 2010; Pfiffner 2014).

**Exploration history**

Exploration for hydrocarbons in the Po Valley started in the first half of the twentieth century (Pieri 1984). Soon after World War II, the investigations progressively covered the NE of the basin, while the use of electric well logs and cores, the development of updated micropaleontological techniques, and, especially, the acquisition of analogue seismic data enabled the recognition and understanding of deeper targets. This resulted in the drilling of the Cavigia 1 well (1944: 1404 m TD bsl (total depth below sea level)), the first gas field discovered by Agip within the Po Valley and the largest in Western Europe at that time. Between 1945 and 1982, the newly acquired digital seismic allowed the very deep horizons to be imaged, also favouring the development of new hypotheses concerning deep lithologies and their associated rock properties. In the 1980s, new methodologies led to the detailed analysis of the seismostratigraphy and the associated structural setting and style of the basin. The integration of well correlations with seismic data and lithofacies interpretations resulted in the construction of the regional base-Pliocene structural map by Pieri & Groppi (1981). From 1973 to 1984, hydrocarbon exploration of the Mesozoic carbonates developed through investigation of both overthrust structures developed during Alpine orogenesis and drilling of Mesozoic structural highs formed by Triassic–Liasic rifting (Bongiorni 1987; Bertello et al. 2010). Both types of targets proved to be successful and led to the discovery of four major hydrocarbon fields, namely the Malossa (gas condensate), Cavone, Gaggiano and Villafortuna (oil) fields. The latter is one of the largest oil fields in continental Europe and has produced 226 MMbbl (million barrels) of light oil to date from a record depth of 6000 m bsl. Today, the Po Valley stands as an underexplored region ready for the next exploration phase, with the help of the exploitation of updated technologies integrated with increased knowledge of the basin geology.

**Hydrocarbon systems and hydrocarbon distribution**

Various petroleum systems have been identified and defined on the basis of drilling, outcrop and systematic analysis of the associated oil and gas types (Riva et al. 1986; Bongiorni 1987; Wygrala 1988; Mattavelli et al. 1993; Lindquist 1999; Bello & Fantoni 2002; Franciosi & Vignolo 2002; Casero 2004; Bertello et al. 2010). The Triassic–Liasic petroleum systems have produced gas, condensate and light oil from deep Mesozoic carbonates (Fig. 1c). The reservoir consists of dolomitized carbonate platform units of middle Triassic–Early Jurassic age, charged by middle–late Triassic carbonate source rocks deposited in intra-platform lagoons and basins. Traps are mostly provided by Mesozoic extensional structures locally inverted during the Cenozoic compression. The Cretaceous–Jurassic pelagic carbonates provide the regional seal. The Villafortuna-Trenta Field (discovered in 1984: light oil; 226 MMbbl of 43°API oil and 93 Bcf (billion cubic feet) of gas produced to date) represents the largest oil accumulation associated with this play (Bello & Fantoni 2002; Bertello et al. 2010). Second-order oil fields in terms of both size and production are the Malossa Field (discovered in 1973: gas and condensate: c. 27 MMbbl and 150 Bcf gas produced) (Errico et al. 1980; Pieri & Groppi 1981; Mattavelli & Margarucci 1992), the Cavone Field (discovered in 1974: 23°API oil; 94.5 MMbbl hydrocarbons initially in place (HClIP) (Nardon et al. 1991) and the Gaggiano Field (discovered in 1982: 36°API oil; 20–30 MMbbl estimated reserves) (Bongiorni 1987; Rigo 1991; Fantoni et al. 2004).

The Oligo-Miocene petroleum system (Fig. 1c) produces thermogenic gas with secondary quantities of oil from the foredeep successions that are detached and thrust over the carbonates and belong to the Northern Apennine belt (Mattavelli & Novelli 1987; Mattavelli et al. 1993; Bertello et al. 2010). The system is composed of thick turbidite sequences that supply both the reservoir and the source and seal elements, and the traps are usually structural, with the Cortemaggiore and Casteggio fields as typical examples of producing fields related to this petroleum system. The Plio-Pleistocene petroleum system contains large volumes of biogenic gas (Fig. 1c), notably at the buried external fronts of the Apennine thrust belt (Mattavelli & Novelli 1987; Mattavelli et al. 1993; Lindquist 1999; Casero 2004; Bertello et al. 2010). The system consists of sand-rich turbidites in which thick-bedded sand lobes and thin-bedded, fine-grained basin plain/lobe fringe deposits are the main reservoir facies associations (Ghielmi et al. 2012). Interbedded clays are both the source rock and the effective top seal. Traps are most commonly structural, yet stratigraphic traps also occur, mainly related to the onlap of turbidite reservoirs onto the flanks of thrust propagation folds or against the foreland ramp. The Settala Field (1977) is a remarkable example of a mixed structural–stratigraphic trap in the Plio-Pleistocene play (Bertello et al. 2010).

The 3D basin model discussed in this paper specifically addresses the burial and temperature history of the thermogenic Mesoozoic petroleum system. The Plio-Pleistocene and Oligo-Miocene systems are not discussed hereinafter.
Methods, input data and modelling assumptions: building and calibrating the thermostructural model of the Po Valley at the Mesozoic carbonate level

Data used for the 3D structural model come from public literature and the archives of the Italian Ministry of Energy (http://unimg.sviluppoeconomico.gov.it; namely, the ViDEPI project). These data include geological cross-sections and well composite logs, as well as geophysical and geological maps. No seismic data have been used during the model building process because: (a) they are poorly distributed across the study area; (b) they are generally low-quality images; and (c) their integration into the model would have required a questionable time–depth conversion, uncertain due to simplifications in the estimated velocity distribution related to the widely varying lithologies in the study area. A full description of the whole dataset, and its distribution across the basin, is provided in Turrini et al. (2014, 2015, 2016). The structural model was built using Midland Valley’s MOVE software (http://www.mve.com/), while progressive refinement of the 3D grids and fault pattern was carried out using IHS’s Kingdom interpretation package (https://www.ihs.com/products/kingdom-seismic-geological-interpretation-software.html).

The resulting Po Valley 3D structural model (Turrini et al. 2014, 2015) consists of 66 faults and five layer grids, namely: the Moho discontinuity, the basement, the near top Triassic, the top Mesozoic Carbonates and the base Pliocene. At all levels within the model, the regional-scale architecture indicates the presence of two crustal domains, a western and an eastern domain separated by the Giudicarie Lineament, a NE–SW–NW–SE extensional fault system in the Liassic, underwent early Cretaceous basinal carbonates in the subsidising Lacchiarella hanging-wall basin. The Gaggiano–Lacchiarella structure (Fig. 5) is a crustal-scale tectonic feature that cuts across the entire Po Valley Basin, and extends towards the Southern Alps to the north and the Northern Apennines to the south (see the Gaggiano location in Fig. 2). This feature has a complex history: it initiated as a north–south-striking, east-dipping extensional fault system in the Liassic, underwent initial inversion in the Oligocene and was weakly reactivated during the Miocene (Fantoni et al. 2004; Turrini et al. 2016). Liassic extension resulted in significant footwall erosion over the crest of the Gaggiano footwall high and in the deposition of an expanded section of deep-water Jurassic and Cretaceous carbonates in the subsiding Lacchiarella hanging-wall basin. Oligocene inversion resulted in approximately no net extension at the top Triassic level across the feature. Inversion and vertical expulsion of the thickened Jurassic–Cretaceous deep-water carbonate sediments, originally deposited in the Lacchiarella hanging-wall basin, resulted in a regional north–south-striking anticline immediately to the east of, and above, the trace of the extensional Liassic fault system (Fig. 5). The structural framework derives from the overprinting of Mesozoic extensional and Tertiary compressional tectonics, as revealed by 2D sections through the model volume (see Fig. 5c–e). Major faults in the region are east-dipping, whereas the associated secondary faults are west-dipping, with the two fault sets bounding the Gaggiano High and the Lacchiarella Basin. The Gaggiano Field (Figs 3a and 5) is located on the west-dipping footwall crest of the north–south Triassic–Liassic extensional fault system (Cassano et al. 1986; Bongiorni 1987; Fantoni et al. 2004; Turrini et al. 2014, 2016). Within the field, the Mesozoic section is extremely thinned by erosion associated with synextensional footwall uplift. Basement is encountered by wells at the exceptionally shallow depths of c. 5 km bsl (Fig. 5c–e). Based on the 3D model reconstruction, the top reservoir at Gaggiano lies just below the top Mesozoic surface, at an average depth of 4.5 km bsl, giving a closure of c. 30 km² and...
Fig. 2. Grid showing the depth to the top Mesozoic carbonates (referenced to mean sea level, contouring every 500 m; bold black lines are major faults at the top of the Mesozoic carbonates); purple lines 'a' and 'b’ show the location of the cross-sections in Figure 3. GFz, Giudicarie fault zone trend line (thick stippled line) separating the eastern domain from the western domain; thin stippled white line marks the area covered by the basin-modelling study described in this paper; bold red line represents the overpressure cell suggested by Chiaramonte & Novelli (1986); Major cities: Mi, Milano; To, Torino; Ge, Genova; Ve, Venezia.
defining a relatively limited four-way dip closure at the crest of the regional footwall (Bongiorni 1987). This trap was formed by Liassic extension and underwent minor rotation during the Cenozoic, along with the deposition of Oligo-Miocene foredeep sediments. The top seal is provided by intra-platform basinal carbonates (Meride Formation), which also form the source rock for the field (Bongiorni 1987; Bertello et al. 2010). Wells drilled on the Lacchiarella inversion structure (Lacchiarella-2 in 1978 and San Genesio in 1994) have encountered significantly increased thicknesses of Jurassic and Cretaceous basinal limestones, confirming the overall tectonostratigraphic model, but have failed to encounter significant hydrocarbons at the Triassic objective levels.

The Malossa Field

The Malossa Field (Figs 3a–6) is located in the western sector of the Milano tectonic arc (see Fig. 2). The field is one of a number of structures that deform the Po Valley Mesozoic foreland and have been buried beneath the Tertiary foredeep wedges to the south of the Southern Alps belt (Errico et al. 1980; Pieri & Groppi 1981; Cassano et al. 1986; Mattavelli & Margarucci 1992; Fantoni & Franciosi 2010; Turrini et al. 2014). The reservoir of the field is provided by fractured late Triassic platform carbonates while the overlying Jurassic–Cretaceous basinal carbonates constitute the seal, along with some further reservoir sections. The source rock has not been proven within the field area. However, analysis of the oil (Mattavelli & Novelli 1987; Mattavelli & Margarucci 1992; Bertello et al. 2010) suggests a late Triassic source rock (Argilliti di Riva di Solto), a lithology which crops out extensively in the Southern Alps, to the north of the Malossa region (Fantoni & Scotti 2003). Stratigraphy from the well information indicates the presence of a Triassic–Liassic high. The trap is provided by a NW–SE-orientated, faulted anticline, plunging towards both the NW and the SE. The associated major thrust is NE dipping and it displaces the structure towards the SW. Minor faults are reported to intersect the fold crest, creating structural compartments within the field (Mattavelli & Margarucci 1992). From the structural model, the average depth to the top Mesozoic structural crest is 5 km bsl, while the field area is c. 15 km² (Fig. 6a). The final age of trap formation is mainly late Miocene, with some minor reactivation during the Plio-Pleistocene (Turrini et al. 2016).

The 3D model (Fig. 6) shows that the Malossa structure was formed by folding and thrusting of the Mesozoic carbonates and the related basement. Sections through the model volume (Fig. 6c–e) confirm that inversion of the Triassic–Liassic extensional basins controls the overall structural style in the region (Cassano et al. 1986; Ravaglia et al. 2006; Fantoni & Franciosi 2010; Masetti et al. 2012) with both the reactivation of Mesozoic extensional faults and the creation of new faults, which locally cut through the pre-existing highs. The Chiari and Belvedere structures to the NE of the Malossa Field are significant, and together with the Lacchiarella structure (Fig. 5) are the main evidence of the basin inversion that took place in the western Po Valley domain (cf. figs 12 and 13 in Turrini et al. 2016).

The key characteristics of these two structures, compared to Malossa, are as follows: (a) the structures are inverted Liassic half-grabens, and the thick (5 km) Mesozoic carbonates are vertically extruded by Miocene inversion (the Malossa structure is essentially
a pre-existing Triassic–Liassic high deformed by Cenozoic thrusting); (b) the Mesozoic faults are reactivated (if the map shown by Mattavelli & Margarucci 1992 is considered correct, it is possible to argue that pre-compressional faults – not represented in the 3D model – are passively displaced by new thrusts in the Malossa structure); (c) some tectonic overthickening of the Jurassic sediments can be interpreted from the public composite log (the Malossa well data do not appear to show any tectonic repetition); (d) the basement is involved in the structuration (as at Malossa); and (e) the age of the present structural geometries is essentially late Miocene with some minor contribution from Pliocene tectonics (as at Malossa).

The Cavone Field

The Cavone Field (Figs 3b–7) is situated on the lateral ramp of a major tectonic arc (i.e. the Ferrara arc) at the buried front of the eastern Northern Apennines (see Fig. 2) (Peri & Gruppi 1981; Cassano et al. 1986; Nardon et al. 1991; Turrini et al. 2014, 2016). The structure is a thrust-related fold where Mesozoic and Tertiary sediments are intensely faulted and fractured (Cassano et al. 1986; Nardon et al. 1991; Carannante et al. 2014). The age of the trap is essentially Late Jurassic carbonate depositional environment (GDE) maps; lateral distribution from gross depositional environment (GDE) maps, thickness and kerogen type as described in the literature; definition of 3D paleo-temperature model by calibration against 1D models for key wells and pseudo-wells; 3D hydrocarbon maturation/ generation/inmigration history modelling across the Po Valley and analysis of kitchen areas associated with key traps.

Table 1. Po Valley 3D basin-modelling workflow and associated working phases

Phase 1 – 1D model building:
- reference well and pseudo-well chrono- and lithostratigraphy, backstripping parameters, thermal parameters, source rock parameters, temperature and maturity data loaded into Genesis (http://www.zetaware.com);
- definition of geological heat flow and overpressure models, primarily based on the available literature;
- collation of information about palaeowater depth and palaeosediment–water interface temperature.

Phase 2 – 1D model calibration and outputs:
- calibration of rock properties and present-day heat flow model against temperature data;
- calibration of back-stripping and heat flow models by forward modelling of thermal maturity and comparison against available maturity data;
- 1D modelling of hydrocarbon generation from key source intervals.

Phase 3 – 3D model building and simulation:
- 3D stratigraphic grids exported from the Kingdom package into the Trinity software, with additional grids generated by interpolating between imported grids as necessary, particularly to define source rock intervals;
- further definition of source intervals within the model, including lateral distribution from gross depositional environment (GDE) maps, thickness and kerogen type as described in the literature;
- definition of 3D paleo-temperature model by calibration against 1D models for key wells and pseudo-wells;
- 3D hydrocarbon maturation/generation/inmigration history modelling across the Po Valley and analysis of kitchen areas associated with key traps.

involvement of the basement particularly unlikely (Cassano et al. 1986; Nardon et al. 1991) unless short-cutting and slicing of the footwall of the foreland unit has occurred (Carannante et al. 2014). The depth to the Cavone culmination from the available public data is c. 3 km bsl to the near top Mesozoic and 4 km bsl to the top Triassic, respectively. According to the reconstructed geometry, the field area would be of the order of 30 km² (Fig. 7a, c, d).

Defining source rock distribution and building gross depositional environment (GDE) maps in the Mesozoic carbonates

Middle and late Triassic intervals (Fig. 8a) are the major source rocks for the deep Mesozoic petroleum system of the Po Valley (Mattavelli & Novelli 1987; Mattavelli et al. 1993; Zappaterra 1994; Lindquist 1999; Katz et al. 2000; Casero 2004; Bertello et al. 2010). A description of the spatial distribution of these source intervals (Fig. 8b, c) and the assignation of the related main parameters describing the hydrocarbon generation potential (e.g. net source thickness, TOC, HI) (Table 1) are, as a consequence, key inputs for the basin modelling. The present section describes how the source model was constrained within the 3D basin model.

The definition of the source rock depositional setting and basin geometry across the Po Valley is a rather difficult task. Indeed: (a) the tectonic history of the basin is complex and polyphased; (b) only a few deep wells have drilled through the Triassic source intervals; and (c) mapping the lateral extent of the source rocks is not easy, given the lack of a clear seismic expression in the basins where the source rocks were deposited. Source rock distribution in the model is consequently described by the construction of gross depositional environment maps (GDE maps) produced for key intervals.

Two loosely defined tectonically controlled megasequences can be identified: (a) a mainly middle Triassic (Anisian–late Carnian) megasequence, associated with extensional–trans tensional tectonics and local volcanism driven by plate-scale wrench movements or aborted rifting; and (b) a mainly late Triassic (late Carnian–early Liassic) megasequence, associated with Tethyan rifting. The middle Triassic megasequence (Fig. 8a) commences with the tectonic segmentation of the widespread epeiric carbonate–evaporitic platform system that dominated in the early Triassic. From the late Anisian onwards, intra-platform basins developed and euxinic conditions occurred periodically. This regional setting resulted in the deposition of organic-rich basinal carbonates over the entire Po Valley realm: the Meride limestone, and the Besano and Gorno formations were deposited in the western Po Valley, whereas the Livinallongo Formation, the bituminous events in the Predil Limestone and the Rio del Lago Formation were deposited in the eastern Po Valley. From the early Carnian onwards, subsidence slowed and platform carbonates prograded across the basins, ending this first phase of deposition of organic-rich facies. The GDE map in Figure 8b shows the interpreted spatial distribution of potential source rock basins for this megasequence; in the western Po Valley, such basins are interpreted to have an approximate north–south orientation, whilst in the eastern Po Valley, the basins are interpreted as orientated NE–SW (Franciosi & Vignolo 2002). In the western Po Valley, two potential source basins are identified: the Anisian–Ladinian Meride-Besano Basin and the Carnian Gorno Basin, situated to the west and east of Milan, respectively. The source potential of the former is confirmed by geochemical correlation with the oils from the Villafortuna-Trecate and Gaggiano fields (Bello & Fantoni 2002). The source rock potential of the Gorno Basin is more speculative: the enrichment of organic matter is reported from outcrops (Assereto et al. 1977; Wygrala 1988; Stefani & Burchell 1990) within sediments deposited in shallow anoxic lagoons developed within a mixed clastic–carbonate depositional system (Gnaccolini & Jadoul 1990). Nevertheless, little direct
evidence exists for hydrocarbons having been generated in the subsurface from that formation. Indeed, extension of this facies southwards, into the subsurface of the Po Valley, is exclusively based on the occurrence of an analogous facies in one of the wells within the Malossa Field. In the central Po Valley, along the buried Ferrara arc (i.e. the buried, external front of the Northern Apennines), the presence of a Mid-Triassic source basin is inferred from the Cavone Field oil-source correlation: this indicates a middle Triassic oil-prone carbonate source rock similar to the Meride Formation of the western Po Valley (Mattavelli & Novelli 1987; Nardon et al. 1991). In the eastern Po Valley and Adriatic foreland, the distribution of potential source basins is taken from Franciosi & Vignolo (2002) with two offshore middle Triassic basins identified, the Ada and Amelia basins, as constrained by 3D seismic. However, the presence of source rock facies remains speculative. Onshore, organic-enriched middle Triassic (Anisian–Carnian) basinal marls and wackestones up to several tens of metres thick are known within the thick basinal successions of the Livinallongo, Predil, Rio del Lago and Durrenstein formations of the SE Alps (Brack & Rieber 1993; Fantoni & Scotti 2003; Keim et al. 2006). Similar facies are encountered in the subsurface of the Po Valley at the Villaverla-1 well: these facies can be interpreted to lie within one of several NE–SW-oriented basins of similar dimensions to those mapped offshore on 3D seismic data (proto-Belluno trough: Masetti et al. 2012).

Extensional tectonics during the middle–late Norian in the Central Southern Alps and in the Carnian Pre-Alps resulted in the progressive segmentation of the widespread Dolomia Principale carbonate platform formed during late Carnian and early Norian quiescence. Extension developed approximately north–south-oriented, intra-platform basins up to several tens of kilometres wide (Jadoul et al. 1992) which expanded as rifting progressed in the Liassic. Drowning of large sectors of the platform led to fully open marine deep-water conditions which were associated with the Tethyan–Ligurian Ocean. Eventually, restricted anoxic conditions developed during the late Triassic. This resulted in the preservation of high levels of organic material within the basinal limestone facies: for example, in the Argilliti di Riva di Solto, Zu and Aralalta formations in the central Po Valley, and the Dolomia di Forni of the eastern Po Valley. The GDE map in Figure 8c shows the interpreted spatial distribution of these potential source basins: the main basin in the western Po Valley is the Riva di Solto Basin of mid–late Norian age. This basin developed in the subsiding hanging wall of the major late Triassic–Liassic Gaggiano-Lacchiarella extensional fault system (Fantoni & Franciosi 2010). Thinner sequences of organic-rich sediments were also deposited in a mid- to outer-ramp setting, in the overlying Rhaetian carbonate ramp represented by the Zu Formation (Stefani & Burchell 1990; Galli et al. 2007). The source potential of these successions is well documented both from outcrop (Jadoul et al. 1992) and geochemical typing of the oils from the Malossa Field data (Mattavelli & Novelli 1987). In the eastern Po Valley, the upper megasequence commences with a widespread late Carnian transgression, resulting in the deposition of the organic-rich dolomites of the Monticello Formation, in an inner-ramp setting. An organic-rich facies, about 60 m thick, ascribed to this interval is reported in the offshore Adriatic foreland at the Amanda-Ibis well (Carulli et al. 1997). As transgression continued into the
Norian, differentiation occurred in areas dominated by the widespread Dolomia Principale Platform, passing laterally into narrow (kilometres to a few tens of kilometres) anoxic basins. An example is the area of the future Belluno Trough where the organic-rich Dolomia di Forni was deposited (Carulli et al. 1997), locally attaining thicknesses of 850 m. Within the Dolomia Principale, anoxic intra-platform lagoons developed locally and these are reported (Carulli et al. 1997) onshore, in the eastern Southern Alps (over 100 m of laminated dolomites and ‘scisti bituminosi’ at Rio Resartico) and in the Adriatic offshore (the Amanda-1bis well).

The GDE maps (Fig. 8b, c) were used to define the lateral source rock distribution within the 3D basin model. Source parameters were then assigned to each polygon. The net thickness of source intervals is poorly constrained: the gross thickness of the source-bearing interval may locally reach 1 km within the major depocentres (Pieri 2001), whilst Fantoni et al. (2002) defined 400 m of gross thickness for the Meride-Besano source interval in the Villafortuna-Trecate Field. On this basis, net source thickness has been assigned with reference to the interpreted GDE, with: (a) long-lived anoxic basins assigned a net source thickness of 50 m; (b) episodically anoxic basins assigned 25 m; and (c) intra-platform/ramp anoxic lagoons assigned 12.5 m.

In general, potential source rocks are carbonate–argillaceous sediments with TOC varying from a maximum of 40% in the Besano Shales to a minimum of 0.10% within the Meride Limestone, with an average of approximately 4% (Novelli et al. 1987; Katz et al. 2000; Fantoni et al. 2002). Kerogen types are dominantly of marine origin, with a secondary component of terrestrial material. For all source rocks within the model, those kerogen types have been parameterized as 90% Type-A kerogen and 10% Type-F kerogen, using default kinetic parameters as defined by Pepper & Corvi (1995a, b) and as shown in Table 2. The only exceptions are the potential source rocks of the Gorno Formation, which are described as dominantly consisting of reworked terrestrial material (Stefani & Burchell 1990) and have consequently been parameterized as 10% Type-A kerogen and 90% Type-F kerogen.

The petroleum potentials derived from these source parameters are reported in Table 2. They appear to be consistent with those reported in the literature: Fantoni et al. (2002) suggested a formation average petroleum potential for the Meride-Besano interval at Villaforunta-Trecate of 21 kg of hydrocarbons per tonne (HC/t) of rock, whilst Bello & Fantoni (2002) indicated a source potential index of 4 t of hydrocarbons per m² (HC/m²) (or 30 million barrels per km² (MMbbl/km²)) for the mid-Triassic petroleum system of the western Po Valley and of 3 t HC/m² (or 22 MMbbl/km²) for the late Triassic petroleum system.

**Model rock physical properties**

The rock properties used as input for modelling include the following: (1) chrono- and lithostratigraphy; (2) surface porosities; (3) compaction coefficients; (4) bulk densities; (5) radiogenic heat generation parameters for each lithology; and (6) thermal conductivities and their temperature dependencies. These
parameters were mainly derived from exploration wells or adjacent outcrop analogues (Berra & Carminati 2010; Pasquale et al. 2011; Pasquale et al. 2012) (Table 3).

The chrono- and lithostratigraphic section used in the 1D modelling was built by assigning the percentages of end-member lithologies present for each stratigraphic unit described (Fig. 9a). Back-stripping and thermal properties were defined based on lithology. For mixed lithologies, properties were derived from the end-member lithologies combined with the relative percentage of each using the appropriate mixing model: simple volumetric weighting was used to calculate surface porosity, compaction coefficient, density, volumetric heat capacity and radioactive heat generation, whilst thermal conductivities were calculated using a geometric mixing law (Pasquale et al. 2011). Temperature dependency of thermal conductivity is incorporated into the model using an approximation to the Sekiguchi Correction (Sekiguchi 1984). A summary of the properties assigned for each end-member lithology is given in Table 3.

**Model pressure in the Mesozoic carbonates**

The Mesozoic carbonates of the western Po Valley are characterized by high overpressures and these represent a significant challenge to deep exploration (Pietro et al. 1979; Vaghi et al. 1980). Early workers argued that formation pressure exerted a significant control on hydrocarbon maturation in the area by illustrating a correlation of the possible overpressures with the difference between observed and theoretically calculated measures of maturity (Chiaramonte & Novelli 1986). While using a global dataset that included data points from the western Po Valley, subsequent investigations highlighted the relationship between vitrinite reflectance and formation overpressure (Carr 1999). This work resulted in a quantitative model based on modifying the Easy%Ro algorithm of Sweeney & Burnham (1990), which is based on the temperature history of a sample, to include an overpressure term. Following the emphasis placed by previous workers in the area on overpressure as a delaying factor on thermal maturity, one of the objectives of the present study was to investigate this effect and, should its importance be confirmed, incorporate it into the 3D basin modelling.

Novelli et al. (1987) briefly reviewed the overpressure distribution in the western portion of the study area. This distribution is characterized by a normally pressured shallow clastic aquifer of Pliocene age and a deep overpressured carbonate aquifer of Triassic age. This latter corresponds to the units that host the Triassic petroleum systems discussed in this paper. The two aquifers are separated by an aquitard consisting of fine-grained clastic rocks of Miocene–Palaeogene age and fine-grained basinal carbonates of Palaeogene–Jurassic age. This aquitard is characterized by a strong pressure ramp connecting the normally pressured shallow aquifer to the overpressured deep carbonate aquifer. These authors interpret overpressures as being due to high sedimentation rates associated with foredeep sedimentation from the Oligocene onwards. Hydraulic isolation of the deep carbonate aquifer occurred during middle–late Miocene times due to Alpine thrusting, resulting in the creation of the deep carbonate pressure cell, in the western Po Valley. Eventually, rapid burial during the Plio-Pleistocene produced the present distribution of overpressure within both the deep carbonate aquifer and the mixed clastic–carbonate aquitard.

**Fig. 6.** The Malossa oil field region (see the location in Figs 1 and 2): (a) top Mesozoic depth grid; (b) 3D structural model of the field and the surrounding structures; and (c)-(e) cross-sections through the 3D model. R/Sr, reservoir and source; Sl, seal. Belvedere well is projected onto section.
In this study, the data and models presented by Novelli et al. (1987) were extended in two ways: (a) by creating 1D pore pressure models for both the aquitard and the deep carbonate aquifer (for key wells), as an input to modelling the thermal maturity of organic matter; and (b) by reviewing the distribution of overpressures within the deep carbonate aquifer against the structure maps from the 3D model, while developing an understanding of the spatial and temporal distribution of these overpressures.

The 1D pore pressure models for individual wells were built in two steps: first, a constant overpressure was estimated for the deep carbonate aquifer, based either on pressure data from the well in question or from data presented by Novelli et al. (1987, their fig. 7); and, secondly, available pressure data (primarily mud-weight data, but with occasional well test or MDT data) in the aquitard were modelled using the Mann & Mackenzie (1990) approach. In this process, the Plio-Pleistocene sedimentation rate was one key input, whilst lithology within the aquitard and top overpressure were other key inputs (Mann & Mackenzie 1990). An example of such a model is shown in Figure 9b for the Belvedere well.

The 3D structural model clearly indicates that the overpressures are confined to a regional-scale anticline developed at the top Triassic level in the western Po Valley (thick red line in Fig. 2), and that this anticline was in place by the end of the Miocene, although it probably formed some time in the Palaeogene (Turin et al. 2016). This anticline is isolated from the normally pressured carbonates of the eastern Po Valley (e.g. the Malpaga-1 well: Novelli et al. 1987) across the Chiari syncline (Fig. 2), which takes the Triassic sediments to a depth of 8 – 8.5 km bsl.

**Model water depths and heat flow**

Palaeowater depths were inferred from: (a) the depositional facies locally defined at the different well locations; and (b) the GDE maps for key intervals (Fig. 8b, c). These depths broadly correlate with those considered by Winterer & Bosellini (1981) for the Mesozoic carbonates, and by Ghielmi et al. (2012) and Di Giulio et al. (2013) for the Cenozoic. Finally, sediment–surface interface and palaeo-temperatures are derived by combining palaeowater surface temperatures, based on the relative latitude of the Po Valley through time, with a discrete water depth–temperature relationship, such as that proposed by Defant (1961).

The heat flow model (Fig. 10) has been defined following a comparative review of published data, primarily from the Southern Alps (Mattavelli & Novelli 1987; Greber et al. 1997; Fantoni & Scotti 2003; Zattin et al. 2006; Scotti & Fantoni 2008; Carminati et al. 2010; Grobe et al. 2015). There is general consensus around two episodes of increased heat flow during the Mesozoic: the first in the middle Triassic, caused by a first pulse of extensional tectonic activity, which resulted in the development of the basins where the middle Triassic source rocks were deposited; and the second during the early Jurassic, associated with the full development of Tethyan rifting. A late Cenozoic reduction in the heat flow trend is observed due to high sedimentation rates and rapid burial in the foredeep, related to the advancing Southern Alps and Northern Apennine fronts. This is consistent with the basin geodynamics and associated tectonostratigraphic evolution of the Po Valley region. The present-day heat flow has been estimated on the regional map of Italy of Fig. 7.
Della Vedova et al. (2001), with corrected well temperature data where available.

**Calibration of the 1D thermal model and assumptions underlying overpressure modelling**

A number of well locations, with available temperature and/or maturity data, were selected for 1D modelling to provide a reasonable geographical spread across the Po Valley region. Maturity data were mainly collected from the literature (particularly Chiaramonte & Novelli 1986; Wygrala 1988; Fantoni & Scotti 2003) with the addition of some proprietary data. Furthermore, some pseudo-wells were constructed to fill in the areas where well data were sparse. The chronostatigraphy and lithostratigraphy for each well were derived from the relevant composite log, with physical properties (porosity, density, thermal conductivity) being assigned based on lithology. Measured temperature data reported on the composite log were corrected to in situ temperature using the approach described by Pasquale et al. (2012). In general, the available maturity data for the Mesozoic carbonates were limited and of poor quality, frequently showing substantial scatter. Much of the data consist of maximum temperature ($T_{\text{max}}$) values from Rock-Eval pyrolysis analysis. These data were converted to vitrinite reflectance (%Ro) equivalent values using the relationship of Jarvie et al. (2001). The satisfactory nature of this relationship in the study area was confirmed at the wells with both $T_{\text{max}}$ and vitrinite reflectance data available.

As a first calibration step, the present-day temperature–depth relationship calculated from the model was compared with the corrected temperature values derived from the composite log. An example is the Belvedere-1 well (Fig. 9c). In general, the match between model and observation was acceptable particularly over the targeted carbonate section. Once a good match was obtained between temperature observations and predictions from the model, maturity profiles were calculated for each well and pseudo-well. Additionally, for wells with maturity data, the calculated profile was compared with observed data. As an example, Figure 9d clearly indicates that the maturity profile calculated using the Easy %Ro algorithm (which uses only the temperature history of each data point: Burnham & Sweeney 1989) for the Belvedere-1 well substantially overpredicts the observed thermal maturity: this is particularly true for the Mesozoic carbonates. In contrast, algorithms that incorporate the overpressure history, in addition to the temperature history, appear to produce a better fit to the observed data, with the PresRo algorithm of Carr (1999) producing very similar results to the alternative $T$–$P$–$Ro$ algorithm of Zou & Peng (2001). It is noteworthy that Carr (1999)
Table 2. Source rock parameters used in the thermal modelling of the Po Valley

<table>
<thead>
<tr>
<th>Source age interval</th>
<th>Domain</th>
<th>Formation(s)</th>
<th>Net thickness (m)</th>
<th>TOC (%)</th>
<th>Kerogen type</th>
<th>Weight (%)</th>
<th>HI</th>
<th>Petroleum potential</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Triassic</td>
<td>Long-lived anoxic basin</td>
<td>Argille di Riva di Solto, Formi</td>
<td>50</td>
<td>4</td>
<td>IIS A</td>
<td>90</td>
<td>550</td>
<td>19.8</td>
</tr>
<tr>
<td></td>
<td>Intra-platform/ramp</td>
<td>Dolomia Principale, Monticello, Calcare di Zu, Scisti Bituminosi</td>
<td>12.5</td>
<td>4</td>
<td>III F</td>
<td>10</td>
<td>160</td>
<td>0.64</td>
</tr>
<tr>
<td>Middle Triassic</td>
<td>Long-lived anoxic basin</td>
<td>Meride, Besano</td>
<td>50</td>
<td>4</td>
<td>IIS A</td>
<td>90</td>
<td>550</td>
<td>19.8</td>
</tr>
<tr>
<td></td>
<td>Episodically anoxic basin</td>
<td>Meride, Livinallongo, Moena, Rio del Lago</td>
<td>25</td>
<td>4</td>
<td>III F</td>
<td>10</td>
<td>160</td>
<td>0.64</td>
</tr>
<tr>
<td>Intra-platform/ramp</td>
<td>Gorno</td>
<td></td>
<td>12.5</td>
<td>4</td>
<td>IIS A</td>
<td>10</td>
<td>550</td>
<td>2.2</td>
</tr>
</tbody>
</table>

Parameters are from published data on the Po Valley Triassic source intervals as reported for the Villafortuna-Trecate and Malossa fields, as well as outcrop analogues. Colours correspond to GDE in Figure 8. Kerogen Types A ('Aquatic, marine, siliceous or carbonate/evaporitic') and F ('Terrigenous, non-marine, wax-poor') are as defined by Pepper & Corvi (1995a, b). These are analogous to Kerogen Type IIS and Kerogen Type III/Type IV, respectively, as defined by Tissot & Welte (1984).

Table 3. Rock properties used in basin modelling of the Po Valley

<table>
<thead>
<tr>
<th>Rock type</th>
<th>Shale</th>
<th>Sandstone</th>
<th>Chalk</th>
<th>Chert/radiolarites</th>
<th>Limestone</th>
<th>Dolomite</th>
<th>Anhydrite</th>
<th>Silt</th>
<th>Marl</th>
<th>Conglomerate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface porosity</td>
<td>0.29(^1)</td>
<td>0.28(^1)</td>
<td>0.70(^2)</td>
<td>0.70(^2)</td>
<td>0.51(^2)</td>
<td>0.30(^2)</td>
<td>0.63(^2)</td>
<td>0.29(^2)</td>
<td>0.50(^2)</td>
<td>0.40(^2)</td>
</tr>
<tr>
<td>Compacton coefficient</td>
<td>0.38(^1)</td>
<td>0.22(^1)</td>
<td>0.71(^2)</td>
<td>0.71(^2)</td>
<td>0.52(^2)</td>
<td>0.22(^2)</td>
<td>0.52(^2)</td>
<td>0.38(^2)</td>
<td>0.54(^2)</td>
<td>0.23(^3)</td>
</tr>
<tr>
<td>Porosity at 3000 m (using Athy equation: (p(x) = \varphi_0 e^{-x}))</td>
<td>0.09(^1)</td>
<td>0.15(^1)</td>
<td>0.08(^2)</td>
<td>0.08(^2)</td>
<td>0.11(^2)</td>
<td>0.16(^2)</td>
<td>0.13(^2)</td>
<td>0.09(^2)</td>
<td>0.10(^2)</td>
<td>0.20(^2)</td>
</tr>
<tr>
<td>Bulk density (kg m(^{-3}))</td>
<td>2720(^2)</td>
<td>2650(^2)</td>
<td>2710(^2)</td>
<td>2650(^2)</td>
<td>2710(^2)</td>
<td>2710(^2)</td>
<td>2270(^2)</td>
<td>2650(^2)</td>
<td>2715(^2)</td>
<td>2650(^2)</td>
</tr>
<tr>
<td>Thermal conductivity (W m(^{-1}) K(^{-1}))</td>
<td>1.62(^3)</td>
<td>3.85(^3)</td>
<td>3.14(^4)</td>
<td>7.14(^4)</td>
<td>3.14(^4)</td>
<td>4.98(^4)</td>
<td>4.76(^4)</td>
<td>3.35(^5)</td>
<td>2.25(^5)</td>
<td>4.18(^6)</td>
</tr>
<tr>
<td>Temperature dependency of thermal conductivity (1/C)</td>
<td>-0.0004(^5)</td>
<td>0.0019(^5)</td>
<td>0.0015(^5)</td>
<td>0.0030(^5)</td>
<td>0.0015(^5)</td>
<td>0.0025(^5)</td>
<td>0.0024(^5)</td>
<td>0.0016(^5)</td>
<td>0.0010(^5)</td>
<td>0.0021(^1)</td>
</tr>
<tr>
<td>Specific heat (J kg(^{-1}) K(^{-1}))</td>
<td>832(^1)</td>
<td>735(^5)</td>
<td>815(^1)</td>
<td>740(^1)</td>
<td>815(^1)</td>
<td>870(^1)</td>
<td>585(^1)</td>
<td>784(^1)</td>
<td>824(^1)</td>
<td>812(^1)</td>
</tr>
<tr>
<td>Specific heat (cal/g/°C)</td>
<td>0.20(^1)</td>
<td>0.18(^1)</td>
<td>0.19(^1)</td>
<td>0.18(^1)</td>
<td>0.19(^1)</td>
<td>0.21(^1)</td>
<td>0.14(^1)</td>
<td>0.19(^1)</td>
<td>0.20(^1)</td>
<td>0.19(^1)</td>
</tr>
<tr>
<td>Radiogenic heat (µW m(^{-3}))</td>
<td>1.33(^6)</td>
<td>1.05(^6)</td>
<td>0.63(^6)</td>
<td>0.43(^6)</td>
<td>0.45(^6)</td>
<td>0.46(^6)</td>
<td>0.09(^7)</td>
<td>1.13(^8)</td>
<td>0.92(^8)</td>
<td>0.90(^8)</td>
</tr>
</tbody>
</table>

Sources: 1Pasquale et al. (2011); 2Berra & Carminati (2010); 3Gretner (1981); 4Middleton (1993); 5Sekiguchi (1984); 6Pasquale et al. (2012); 7Waples & Waples (2004). Where available, local rock properties are used (Berra & Carminati 2010; Pasquale et al. 2011, 2012); other values are from global averages (Gretner 1981; Middleton 1993; Waples & Waples 2004).
Fig. 9. Synthetic well logs for the Belvedere 1 well (depth is in metres): (a) chrono- and lithostratigraphy; (b) formation pressure model showing the significant increase in overpressure below 2000 m through the Tertiary foredeep clastics and basinal carbonates into the highly overpressured deep carbonate aquifer consisting of Liassic and Triassic platform limestones and dolomites; (c) temperature model showing good correspondence between corrected well-temperature measurements and the prediction from the basin model. The average temperature–depth trend for the western Po Valley from Pasquale et al. (2012) together with the observed range is also shown; and (d) thermal maturity model showing the match of various models to the dataset from Chiaramonte & Novelli (1986).
incorporates overpressure effects into the Easy%Ro model by introducing a pressure-based modification to the frequency factor, whilst Zou & Peng (2001) introduced an overpressure based modification to the activation energies. For the purposes of this modelling exercise, it was assumed that pressures were hydrostatic up to the end Miocene isolation of the deep carbonate aquifer in the western Po Valley. From the end of the Miocene onwards, it was assumed that overpressures increased linearly with time up to the present-day values modelled for a particular interval. As for Belvedere 1, other wells included in the dataset showed similar results, with an improved fit to observed maturity data from models incorporating overpressure and overprediction of maturity using Easy%Ro. Of particular note at the Belvedere-1 well is the way in which the results of the overpressure algorithms converge with the Easy%Ro model below 7500 m TVDs (true vertical depth subsen) (Fig. 9d). This is likely to be due to peak maturity deep within the carbonate section having been achieved during the Liassic rift event, long before significant overpressure entered the system. Such an early maturity was a consequence of the thick synrift section deposited at this location, combined with elevated heat flows. Notwithstanding the relatively poor quality and scattered nature of the maturity data, this analysis would appear to support the inference that overpressure has delayed the thermal maturity of the Triassic source rocks in parts of the Po Valley as suggested by Chiaramonte & Novelli (1986) and Carr (1999).

The Genesis and Trinity 3D modelling software from Zetaware Inc. used in this study does not incorporate algorithms that include the overpressure effect. The most appropriate modelling strategy was therefore to approximate the overpressure effect in the software by applying a reduced heat flow, given that overpressure appears to act to delay maturation (Carr 1999). Figure 9d shows that the maturity profiles calculated for the Belvedere-1 well using the overpressure algorithms are approximated by a temperature-only maturity model using a heat flow that is 15 W m$^{-2}$ lower than the currently observed heat flow at this location. Hence, to replicate the overpressure history in the basin, the reduced heat flow model was built to equal the geological heat flow up to the end of the Miocene. From that moment, the heat flow was varied linearly to reach a present-day value that is 15 mW m$^{-2}$ lower than the observed present-day heat flow. Similar results were obtained for other wells in the dataset. This analysis was also repeated for a number of pseudo-well data points covering the depth range of the Triassic source rocks within the model overpressure cell. This operation confirms that a reduced heat flow model satisfactorily replicates the maturity trends generated by the overpressure model.

Modelling results

1D thermal model and hydrocarbon generation

The results from 1D modelling for well and pseudo-well locations in the western, central, east-central and eastern Po Valley are summarized in Figure 11. For the western and central Po Valley, two sets of results are provided, one based on the actual geological heat flow and one which considers the effect of overpressure through the application of the reduced heat flow model from end Miocene times. In the western Po Valley, west of Milan (Fig. 11a), the Triassic source intervals reached maturity during the Miocene as a result of burial beneath the thick Alpine foredeep sediments. These source rocks are currently in the late oil window. In contrast, in the central Po Valley east of Milan (Fig. 11b), Triassic source rocks started generating hydrocarbons during the Jurassic, with renewed generation in the Miocene, and are currently in the late oil to gas windows. This generation process is probably due to the increased thickening of synrift Liassic carbonates in the hanging wall of the Gaggiano-Lacchiarella fault system, combined with high synrift heat flows. For both the western and central Po Valley well locations, the reduced heat flow/overpressure model shows lower maturity, all through the Plio-Pleistocene. In the western Po Valley, this equates to the difference between middle oil maturity (%Ro value of c. 0.8) and wet gas maturity (%Ro value of c. 1.3).

Over most of the eastern Po Valley, middle Triassic source rocks attained early maturity during the Jurassic (Fig. 11c) due to thick carbonate deposition and high heat flows, with only minor increases in maturity to the present day as a result of lower heat flow and/or a low sedimentary depositional rate. During the same time interval, late Triassic source rocks remained immature to very early mature (Fig. 11c). Figure 11d shows the 1D model for part of the Trento Platform in the eastern Po Valley where sedimentation rates remained particularly low. In this location, only limited generation potential is envisaged, with the early oil window being reached by the middle Triassic source rocks in the late Miocene–Recent, whilst late Triassic source rocks are essentially immature at the present day.

3D thermal model and hydrocarbon generation

Results from 1D modelling (see above) and GDE maps have been integrated with the 3D structural model to create a 3D thermal model of the entire Po Valley foreland basin. Using the 1D well models as anchor points, two thermal histories were created and calibrated to best represent the thermal histories of the middle and late Triassic
Fig. 11. 1D transformation ratio (TR) maturity histories for four wells from the Po Valley based on initial source rock parameters outlined in Table 1 (the TR scale is 0–100): (a) Cerano-1 from the western Po Valley; (b) Belvedere-1 from the central Po Valley; (c) a pseudo-well from the east-central Po Valley; and (d) Ballan-1 from the eastern Po Valley (see Fig. 2 for the well locations). Vitrinite reflectance maturities are shown as blue lines (note that for wells in a & b, two histories are shown for the last 10 myr: one based on the geological heat flow and one based on reduced heat flow from end Miocene times to replicate the effect of overpressure; wells in c & d lie outside of the overpressure cell; see the text for explanations).
source intervals, one based on the actual geological heat flow model and one based on the reduced heat flow to replicate the effect of overpressure. In particular, the reduced heat flow associated with the overpressure model is confined to the area of the regional-scale anticline at the top Triassic level that contains the overpressure cell, as shown in Figure 2. Outside this area, the two heat flow models are equal.

The progressive change in transformation ratio (TR) through time across the Po Valley for the middle and late Triassic source intervals from the Mesozoic to the end Miocene is illustrated in Figure 12. For middle Triassic source rocks, early oil maturity is attained during the Jurassic to the east of the Gaggiano Lacchiarella fault system and in most of the eastern Po Valley, whilst to the west, maturity remains low (Fig. 12a). This clearly fits the 1D modelling scenarios and confirms the results presented by Novelli et al. (1987).

The maturity pattern is attributed to high synrift heat flows associated with Liassic rifting, combined with the deposition of (a) thick sequences of basinal limestones in the hanging wall of the Gaggiano Lacchiarella fault system, (b) thick shallow-marine carbonate deposits in the area of the Trento Platform (Fig. 2) and (c) thinner basinal sequences to the west (footwall) of the Gaggiano Lacchiarella tectonic trend. Through the Cretaceous, only small increases in maturity are observed due to low sedimentation rates in a deep-water basinal setting. During this period, heat flows returned to typical passive-margin setting levels (Fig. 12b) (Fantoni & Scotti 2003). Remarkably in Jurassic and Cretaceous times, the late Triassic source rocks remain immature, except in the vicinity of locally thick carbonate deposits, particularly in the central and NW Po Valley (Fig. 12d, e).

During the early Tertiary and up to the end of the Miocene, the enhanced clastic influx from the Southern Alpine and Northern Apennines thrust belts increased burial of both Triassic source intervals with further increases in maturity. Locally, where sedimentation rates were highest, such as in portions of the Southern Alpine foredeep, this resulted in the completion of the kerogen transformation process (Fig. 12c–f). Notwithstanding this,

![Fig. 12](Figures/12.png)

Fig. 12. Transformation ratio (TR) maturity maps (the TR scale is 0–1) for the middle Triassic (a)–(e) and the late Triassic (d)–(f) source intervals, for end Jurassic (a & d), end-Cretaceous (b & e) and end Miocene (c & f) times. As the onset of overpressure within the carbonate sequences is interpreted to occur at the end Miocene, there is no difference between the maturity levels associated with the geological heat flow and the overpressure models for this time interval.
the Liassic structural grain continued to exert an influence on maturity patterns with much of the Gaggiano footwall and Trento Platform constantly exhibiting low maturities.

In middle–late Miocene times, the deep carbonate aquifer in the western Po Valley became isolated and the Triassic source intervals started to experience overpressure. Figure 13 compares the present-day TR distribution for the actual geological heat flow and reduced heat flow/overpressure models. The high Plio-Pleistocene sedimentation rate resulted in increased maturity throughout the Po Valley; however, as expected, within the western Po Valley overpressure cell, the increase in maturity is substantially less for the overpressure model than for the geological heat flow model (cf. Fig. 13a–c and 13b–d). This effect is particularly evident over the crest of the Gaggiano footwall: the area shown in blue at the end Miocene for both middle and late Triassic intervals (Fig. 12c, f), corresponding to a TR of less than 10%, has completely disappeared at present day for the geological heat flow model (Fig. 13a–c), whilst for the overpressure model narrow belts with low TR remain over the crest of the footwall region (Fig. 13b–d).

Remarkably, both models show hydrocarbon generation occurring in two phases (Figs 11, 12, 13 and 14): a Jurassic phase and an Alpine Tertiary phase, the latter starting in the Oligocene but occurring mainly during the last 5–10 myr, in agreement with earlier findings (Mattavelli & Novelli 1987; Novelli et al. 1987; Mattavelli et al. 1993; Lindquist 1999; Bertello et al. 2010).

Discussion

Overall validity of the thermostructural modelling approach and choice of the better model

3D charge modelling was carried out for a number of structures within the western Po Valley overpressure cell in order to compare model predictions with observed hydrocarbon distribution and properties. Charge modelling was performed using the simple kinetic methodology described in Pepper & Corvi (1995a, b) and Pepper & Dodd (1995) as implemented in the Trinity Basin Modelling software. Source rock kerogen types and initial HIs and TOCs values are shown in Table 2. For each structure, kitchen areas were defined as the areas of the present-day top Triassic depth map over which buoyancy forces would drain migrated hydrocarbons towards the relevant structural culmination. These areas were then further refined by superimposing the source rock polygons from the GDE maps. Finally, the charge volumes for the various traps were then limited to those available after the critical moment: that is, the time at which the trap formed or the seal became able to retain a hydrocarbon column (Fig. 14). The model also incorporates the effect of migration losses along the path to the trap, with considered loss of 0.075 MMbbl/km², derived using the methodology proposed by Mackenzie & Quigley (1988) with a bed thickness of 500 m and an average porosity of 1.5%. Reservoir and top-seal parameters are defined in order to allow the basin model to calculate volumes trapped in each structure. Here, a single late Triassic reservoir was modelled as a 250 m-thick, 100% net-to-gross slab with an average porosity of 3% (see Bello & Fantoni 2002 for comparison). Top-seal capacity was modelled as 300 psi using simple capillary seal models for pelagic carbonates. The basin model has been re-run, and the following predicted parameters were extracted: volume of charge available from the relevant kitchen area since the critical moment, trapped hydrocarbon volume and gas/oil ratio (GOR) of the trapped fluids.

These predicted parameters compare well with estimates of the initially in-place hydrocarbon volume (HClIP) at each trap and for the GOR of the fluids present in the three main discoveries in the western Po Valley (Fig. 15): to a first order, both the actual geological heat flow and the reduced heat flow/overpressure models.
Fig. 14. Charge timing v. trap formation in the western Po Valley based on the preferred overpressure model (see the text for the discussion). Vertical orange arrows indicate the presumed critical moment for each of the traps (i.e. the time at which the trap formed or the seal became able to retain a hydrocarbon column). mmboe, million barrels of oil equivalent.
Fig. 15. Model evaluation: (a) cross-plot of observed in-place volumes for main traps v. available charge from the kitchen area since the critical moment predicted by the models; (b) cross-plot of observed volumes in-place for main traps v. predicted trapped volumes from the models; (c) cross-plot of observed GOR v. predicted GOR from the models. Red data points and regression lines are for the geological heat flow model; blue data points and regression lines are for the overpressure model. In (b), regression lines have been fitted to the dataset excluding the Gaggiano outlier. In all plots, the black line corresponds to a perfect match between observation and model. bbl, barrels; mmboe, million barrels of oil equivalent; scf, standard cubic feet.
replicate accurately the overall distribution and phase of hydrocarbons and predict significant discoveries at Villafortuna-Trecate and Malossa and a smaller discovery at Gaggiano. They also predict a rich petroleum system with significant volumes of hydrocarbons spilled from traps that have been breached, bypassed and/or overfilled. This is evident at Gaggiano where the two models equally calculate small trapped volumes due to the size of the trap. Indeed, being located at the crest of a regional high (see Figs 2 and 3a), the Gaggiano trap appears to be linked to an extensive kitchen area, which, since the Mid-Miocene critical moment, has generated charge volumes 25–50 times larger than the trapped volumes. Finally, the two models predict liquid hydrocarbons with moderate-to-low GOR at Villafortuna-Trecate and Gaggiano, whilst high GOR fluids are predicted at Malossa.

As a result, despite the relative simplicity of the modelling approach adopted and uncertainties regarding source rock distribution, our 3D thermostructural modelling provides for the first time a consistent integration of the 3D structures with their thermal histories and reliably simulates the related hydrocarbon maturation/generation process across the entire Po Valley Basin.

In detail, however, the reduced heat flow/overpressure model better matches the observed data than the actual geological heat flow model. In this respect, Figure 15a compares calculated trap HCIIP volumes with the predicted charge available from the kitchen area since the critical moment. The graph shows that predictions from the overpressure model (excluding Gaggiano) correlate better with trap HCIIP values than those from the actual geological heat flow model. Also, the overpressure model can successfully explain the failures in the inversion traps in the Lacchiarella hanging wall (Lacchiarella and San Genesio) and the deep traps east of Malossa (Chiari, Belvedere). Conversely, the actual geological heat flow model predicts significant volumes in several of these traps. Furthermore, charge volumes available to the trap are closer to HCIIP volumes for the overpressure model than for the actual geological heat flow model. This implies that smaller volumes are spilled to shallower traps and/or stratigraphic levels. Given the little evidence for large spilled volumes in the Po Valley, the prediction of smaller excess volumes favours the overpressure model.

Figure 15b shows how predicted trap volumes from the basin models compare with the calculated trap HCIIP volumes. Given that traps are generally oversupplied with hydrocarbons in both models, there is relatively little difference in the performance of the two models. However, it is of note that Malossa volumes are matched better by the overpressure model as there is a charge limitation on predicted volumes in the trap; the actual geological heat flow model predicts larger volumes with the trap being oversupplied and excess volumes spilled. Finally, Figure 15c shows that the overpressure model more successfully predicts fluid phase than the actual geological heat flow model, which predicts higher maturity fluids with higher GORs than observed for all three of the main discoveries.

We therefore conclude that overpressure as simulated by a reduced heat flow is a viable and valid mechanism that has probably significantly delayed hydrocarbon maturation in the western Po Valley, as proposed by earlier authors (Chiaramonte & Novelli 1986; Carr 1999).

Uncertainties on the modelling results and sensitivity

Structural model uncertainties

The Po Valley 3D structural model (Turrini et al. 2014) defines the present-day configuration and geometrical framework of the basin. Although a regional-scale kinematic restoration to pre-Alpine and/or Mesozoic position has been recently attempted (Turrini et al. 2016), the chosen modelling approach applied here to the evolution of the Mesozoic petroleum system is a conventional one. Although a 2D kinematic approach would have been a more accurate methodology for modelling such a complex petroleum system (Gusterhuber et al. 2014; Neumaier et al. 2014), simple vertical back-stripping was carried out to describe the tectonostratigraphic evolution of the basin. Despite this simplification, we believe the modelling results are reasonable due to the following considerations.

The model has been restricted to the foreland domain, characterized by low deformation and in which vertical displacements are more significant than horizontal ones (Cassano et al. 1986; Turrini et al. 2014). Locally, thrust faults can create a late tectonic overthickening of the thrust section, particularly where a hanging-wall ramp is juxtaposed with a footwall ramp. An example is provided by the Medolo Formation in the Belvedere well, where an estimated 500 m of tectonic thickening occurs on a Miocene thrust fault. This is incorporated into the model as stratigraphic thickening of the Medolo sediments and contributes to the high TR in the vicinity of the Belvedere well shown at end Jurassic times (Fig. 12a, d). However, sensitivity modelling indicates that the effect is minor and local, given the relatively small scale of the thrusting involved, and does not impact the validity of the regional results presented.

The vertical back-stripping approach used approximately describes the recent evolution of the system, and covers the bulk of hydrocarbons generated during the Alpine phase. The model will not adequately describe the generation and expulsion of hydrocarbons during the earlier Jurassic phase as trap distribution and geometry were substantially different during this phase. However, the effective charge in both models has been limited to a post-critical moment that took place some time in the Miocene. Consequently, hydrocarbons generated earlier are lost to the system and are deemed to have leaked to the surface. Therefore, the lack of structural restoration does not impact the results, although any possible re-migration from reactivated Mesozoic traps has not been considered.

A further simplification in the model is that all surfaces other than the base Pliocene surface have been modelled as conformities. A number of erosional unconformities earlier in the Tertiary have been neglected due to insufficient data to simulate these at the regional scale of the model. The literature on the region (Pieri & Groppi 1981; Cassano et al. 1986; Ghielmi et al. 2012; Rossi et al. 2015) suggests that: (a) erosion of Mesozoic sediments was essentially restricted to locally uplifted areas, such as the synrift footwall erosion experienced over the crest of the Gaggiano footwall; and (b) erosion of Tertiary deposits associated with intra-Tertiary unconformities is of the order of a few hundred metres. Consequently, given the limited pre-Pliocene erosion and high Pliocene–Pleistocene sedimentation rates, it is likely that Mesozoic source rocks are at maximum depth of burial and peak thermal maturity at the present day across the vast majority of the basin (Ghielmi et al. 2012; Rossi et al. 2015). Given the limited and local nature of the pre-Pliocene unconformities, it is considered unlikely that their absence from the model significantly affects results, although it may result in some local errors in the maturation history.

Petroleum systems uncertainties

The main uncertainty pertaining to petroleum systems consists of the source rock distribution (position and areal extent of the source polygons of Fig. 8b, c) defined on the basis of the GDE maps. A second major uncertainty refers to the assigned net source rock thicknesses, essentially due to the paucity of the available input data. Indeed, the models mainly rely on outcrop information from the Southern Alps and it should be noted that the South Alpine Front, which separates the outcrops from the subsurface of the Po
Valley, is a Tertiary feature with an estimated 50–70 km of shortening (e.g. Handy et al. 2014). In this framework, considerable uncertainty exists in correlating from the outcrop to the subsurface. Furthermore, the source rock distribution defined here includes a number of postulated source basins, particularly in the eastern Po Valley and the Adriatic offshore.

Another potential issue arises in the interpretation of the unsuccessful wells in the western Po Valley. The ability to explain these failures as due to a lack of access to recent charge was used as a reason for preferring the reduced heat flow/overpressure model to the actual geological heat flow model (the latter predicting the availability of significant recent charge volumes to these traps). Clearly, there is a range of other potential failure mechanisms unrelated to source rock that could explain these well results.

**Sensitivity to thermal and burial history parameters**

The basin modelling presented here derives from a long and continuous analysis of sensitivities for the many parameters which control the burial and thermal history of the Po Valley region.

Heat flow based on data from the available literature (see Fig. 10) was chosen as the key element to replicate the overpressure effect. Reducing the heat flow is a straightforward method to control the vitrinite maturation progression around the basin. In addition, using heat flow as a key controlling factor for hydrocarbon maturation can be used as a stand-alone tool that does not directly impact the various parameters which affect the simulation process (e.g. rock properties, burial history, source distribution). Quality control (QC) on the heat flow history was concentrated on both past and present history to best match the vitrinite profile available at selected well locations in the Po Valley. In particular, in order to build the reduced heat flow/overpressure model, particular attention has been paid to the reconstruction of the Miocene–Plio-Pleistocene curve segment. This needed to be viable with respect to the tectonostratigraphic history of the basin where rapid sedimentation of the clastic succession was associated with localized overpressure build-up in the Mesozoic carbonates. The radiogenic heat flow component possibly derived from mineral associations of the Tertiary sediment has also been evaluated, although it was finally considered irrelevant to the basin model results.

Notwithstanding the key role of the Po Valley heat flow on the study objectives, all of the basin model parameters (see Table 2) have been progressively evaluated and implemented from the initial Genesis/Trinity software standard values. Again, the primary aim was to refine the match with the available maturity data while keeping a present-day heat flow consistent with the published one. In particular: (a) lithologies have been refined on the basis of a careful analysis of the well logs; (b) matrix thermal conductivity of the sediments, especially for shales and sandstones, has been reviewed in the light of the available literature; (c) for specific rock types, such as silts and conglomerates, surface porosity, compaction coefficient, porosity and bulk density have been adjusted using literature data while iteratively validating the model constraints (i.e. well temperatures and vitrinite profiles); and (d) porosity in the Mesozoic carbonates was also validated against the field values as it was considered the main variable in the computation of migration losses in the model. Observed hydrocarbon production analysis.

Further sensitivity tests were performed on progressive sea-level palaeodepth variations, an important control on sea-level temperature at the different stages of the burial–thermal history. Indeed, almost all of the decrease in water–sediment interface temperature occurs in the first 100 m, so that anomalously shallow palaeodepth estimates can cause 10°C excess temperature at the source rock level through part of the geological burial history. This would then require an unrealistic reduction in the heat flow in order to match the vitrinite data constraining the basin model.

Finally, the properties and parameters that have been used and progressively implemented during the model building are strictly interrelated. Sensitivity analyses demonstrated how changing one parameter often results in a compensatory change to another parameter. Their implementation, coupled with heat flow adjustment, had a significant impact on the final model results.

**Implications for the thermostructural evolution of the Po Basin, and hydrocarbon generation and prospectivity**

The 3D basin model of the Po Valley presented in this paper provides important insights into the geometry and structural evolution of hydrocarbon-bearing traps, and into the generation and migration of hydrocarbons into these traps.

The model confirms earlier studies (Mattavelli & Novelli 1987; Novelli et al. 1987; Mattavelli et al. 1993; Lindquist 1999; Bertello et al. 2010) and shows that hydrocarbon generation is likely to have occurred in two phases: a Jurassic phase and an Alpine Tertiary phase, the latter occurring mainly during the last 5–10 myr. Our results emphasize the impact that Mesozoic and Tertiary Alpine tectonics had on the development of a successful petroleum system in the Po Valley. The Mesozoic extensional phase controlled reservoir and source distribution, trap formation (e.g. the Gaggiano oil field), and the early phases of hydrocarbon maturation in subsiding half-graben associated with high heat flows and substantial synrift to early post-rift sediment accumulation. The Tertiary compressional phase controlled trap formation, either by generating new traps (the Cavone oil field) or by reactivating older ones inherited from the Mesozoic extensional phase (the Villafortuna-Trecate and Malossa oil fields). Clearly, regional hydrocarbon maturation and expulsion/migration are related to rapid foredeep burial ahead of the evolving Southern Alpine and Northern Apenninic thrust belts.

From a hydrocarbon exploration point of view, the timing of hydrocarbon maturation is favourable for exploration in the western Po Valley. Trap formation is likely to have occurred during the Oligocene–late Miocene, along with significant post-Miocene hydrocarbon generation and expulsion (migration?). In contrast, in the eastern Po Valley, timing is less favourable as traps (Plio-Pleistocene in age) tend to either post-date the main hydrocarbon generation phase or they formed when generation was not advanced enough for migration to occur, or for traps to be filled.

**Conclusions**

Using the recent Po Valley 3D structural model as an input for basin modelling, the approach presented in this contribution provides for the first time a unique integration of the 3D structures with their thermal history and the related hydrocarbon maturation/generation process across the entire Po Valley Basin. When compared with the observed distribution of hydrocarbons, our basin modelling results suggest that, at the regional scale, both maturity models (actual geological heat flow model and reduced heat flow/overpressure model designed to simulate the delaying effect of overpressure on hydrocarbon generation) appear consistent with the observed hydrocarbon distribution. In detail, however, the overpressure model (a) provides an improved match to observed maturity data, (b) provides a better fit between calculated trap HCIP volumes and predicted charge available from the kitchen area since the critical moment, and (c) predicts the hydrocarbon phase (as measured by GOR) more accurately than the geological heat flow model. However, caution should be applied to the different variables and uncertainties that pertain to the accumulation process (i.e. source rock net pay, expelled v. unmovable hydrocarbons, heterogeneity in the TOC content of the source intervals, reservoir net volume and associated heterogeneity, and quantitative estimates of migration.
losses). The modelling results confirm that the delaying effect of overpressure is an important factor to be taken into account in predictions of hydrocarbon maturation and generation.

The study also confirms the impact that Mesozoic and Tertiary Alpine tectonics had on the development of a successful petroleum system in the Po Valley. The Mesozoic extensional phase controlled reservoir and source distribution, trap formation, and the early phases of hydrocarbon maturation in subsiding half-graben associated with high heat flows and substantial synrift to early post-rift sediment accumulation. The Tertiary compressional phase controlled trap formation, either by generating new traps or by reactivating older ones inherited from the Mesozoic extension.

This study demonstrates the utility and applicability of a consistent integrated 3D model of the thermostructural history of sedimentary basins to constrain the geometry and structural evolution of hydrocarbon-bearing traps, as well as the generation and migration of hydrocarbons into these traps.

Acknowledgements

Roberto Fantoni from ENI S.p.a. is kindly thanked for discussions about some parts of the manuscript. We thank Jo Prigmore, Tim Diggs and Ozkan Hozar for their constructive comments on the manuscript.

References


