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QUATERNARY STUDIES AS A TOOL TO VALIDATE SEISMIC HAZARD POTENTIAL OF TECTONIC STRUCTURES: THE CASE OF THE MONFERRATO THRUST FRONT (VERCELLI PLAIN, NW ITALY)

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ABSTRACT: This paper proposes the study of the Quaternary geological evolution of the Vercelli Plain (Piedmont, NW Italy) with the aim of validating the reliability of assumptions about seismic hazard of tectonic structures based on seismotectonic data obtained from studies at regional scale. In particular, the Quaternary evolution is interpreted in order to verify and date the tectonic activity of the Monferrato thrust front, i.e., the westernmost arch of the buried northern Apennines front.

In reference to the main purpose of the work, it is observed that only some stretches of the buried front (Lucedio and Cavourrina fault and, very likely, the flexures of Crescentino and Morano Po) were active between 870 and 400 ka BP.

After 400 ka BP, near Crescentino and Trino, some structures transversal to the Monferrato front were probably activated, inducing the uplift of N-S elongated areas. The uplift affected both the northern Monferrato slope and the areas of the plain located south and north of the thrust front. The uplifted Trino area was limited to the east by the Salera Line. There are no indications of structures (fault or flexure) to the west of the Trino area or which form the limits of the uplifted Crescentino area, although their presence could be hypothesized.

The Salera Line is the most important identified structure; it is very likely a complex-kinematics fault that was active from the Pliocene to the Upper Pleistocene and the Holocene. It continues to the south inside the hill areas, for a length of about 20 km.

Therefore, the recent tectonic movements are not associated with the Monferrato thrust front, as claimed by other authors, but rather with N-S structures which are transversal to the front. Although elements to evaluate the seismic hazard possibly associated with these structures are not available, it is unlikely that the seismicity alone of the easternmost Apennine fronts (Emilia and Ferrara Folds) can provide useful information to assess the seismic hazard of the Vercelli Plain and of Northern Monferrato.

KEYWORDS: quaternary evolution, buried tectonic structures, climatic impact, climate and tectonic interaction, river diversions, Vercelli Plain, Monferrato hills.

1. INTRODUCTION

A detailed study of the Quaternary evolution of the Vercelli Plain (Northwestern Italy) is used here as a tool to assess the reliability of previous assumptions based on seismic hazard which were generally based on the analysis of historical seismicity, on the evolution of the drainage network, on the presence of Quaternary sediments displaced by faults, and on GPS measurements.

Data collected and presented in this paper are interpreted in order to verify and date the tectonic activity of the buried thrust front of this sector of the Apennines. As Quaternary sediments of the Vercelli Plain bury the Monferrato thrust front (i.e., the westernmost of the three major buried arches of the Northern Apennines; Fig. 1A) the detailed geological and geomorphological study of the plain provides a reliable tool for understanding the possible activity of the front.

In fact, according to several Authors (e.g. Costa, 2003; Dela Pierre et al. 2003b; Galadini et al., 2012) the activity of the frontal thrust of Monferrato ended in the early Pleistocene while, further to the East, the Emilian and Ferrara fronts are still active. This is testified also by historical and instrumental seismicity, which is totally absent in the Monferrato front, and increases toward the Emilian folds (Galli, 2005), reaching a maximum in the Ferrara Folds (Galli et al., 2012; Burrato et al., 2012a).

On the other hand, the results of other seismotectonic research, conducted at the scale of the whole Po Valley, led to the hypothesis that the Monferrato front may be seismically active, even if characterized by earthquakes with long return periods. Both Bonadeo et al. (2010) and Michetti et al. (2012), suggest that the level of seismic activity of the Monferrato front could be similar to the two other Northern Apennine fronts. Indeed, according to Michetti et al. (2012), the Monferrato front might be able to generate earthquakes of Mw 6.

In the past, the presence of the Trino isolated ridge (known as RIT in the literature), lying above the buried Monferrato front, had already stimulated research by the Working Group on the Po Valley Quaternary (Gruppo di Studio del Quaternario Padano: GSQP, 1976) which established that the area could have been affected by Middle Pleistocene tectonic activity. Moreover, geodetic data reported in Arca & Beretta (1985) provide information on some aseismic deformations that occurred along a section which crosses the SE portion of the area of this study during the period between 1897 and 1957.

Thus, the conflicting interpretation of tectonic activity and seismic hazard, and the presence of aseismic deformations, require detailed study of the Quaternary geological evolution to determine which deformations were recorded by sediments and landforms, and thus which structures are likely to be active.

In this paper, the evaluation of the deformation of Quaternary sediments, including those dated as the last 2-3 ka, suggests the activity of tectonic structures which only partially match those hypothesized in the literature.

To assess adequately the tectonic evolution of this area, this study takes into account the interaction between sedimentary and erosive phases driven by climate and those probably influenced by tectonics, starting from the upper Lower Pleistocene.

2. STUDY AREA

The Vercelli plain lies between the river Dora Baltea and the Ivrea morainic amphitheatre to the west, the river Po and the Monferrato hills to the south, the river Sesia to the east and the rivers Elvo and Cervo to the north (Fig. 2A). The plain is mainly formed by Plio-Pleistocene sediments, but in its southern part a thin Quaternary sedimentary cover lies on marine sediments. The Tertiary marine sediments, outcropping in the Po riverbed, are folded and heavily deformed by the activity of the northernmost Monferrato thrust front (SGd'I, 1969a; Dela Pierre et al., 2003a).

A large part of the Vercelli plain is in the Sesia River catchment whereas only the western and southern portions of the plain are in the catchments of the rivers Dora Baltea and Po (Fig. 2B).

It should be noted than the Po catchment, which includes a large area of the western Alps and the northern slope of the Torino and Monferrato hills, before receiving the Stura di Monferrato stream, narrows to 1.5-2.5 km (Fig. 2B). In the Monferrato hills the divide of the small catchments that drain toward the Po migrates abruptly towards the north and remains very close to the plain down to the confluence with the Stura stream. Starting from the area of Gabiano, the bed of the Po is always very close to the hill slope and erodes it. The narrowing of the Po catchment in the southern Vercelli plain and in the northern Monferrato slope is an apparent anomaly compared to the westernmost areas.

The Po, which is the main watercourse in the study area, during the Quaternary was mostly flowing south of the Monferrato hills. It migrated north of the hills only after a diversion that took place during the Late Pleistocene (Carraro, 1976; Carraro et al.,1980; 1995).

Before the diversion, the rivers Sesia and Dora Baltea flowed into a river, which from now on will be referred to as the DROS, formed by the confluence of the rivers Dora Riparia, Orco, Stura and other smaller streams, whose basin was in the north-western part of the Alps.

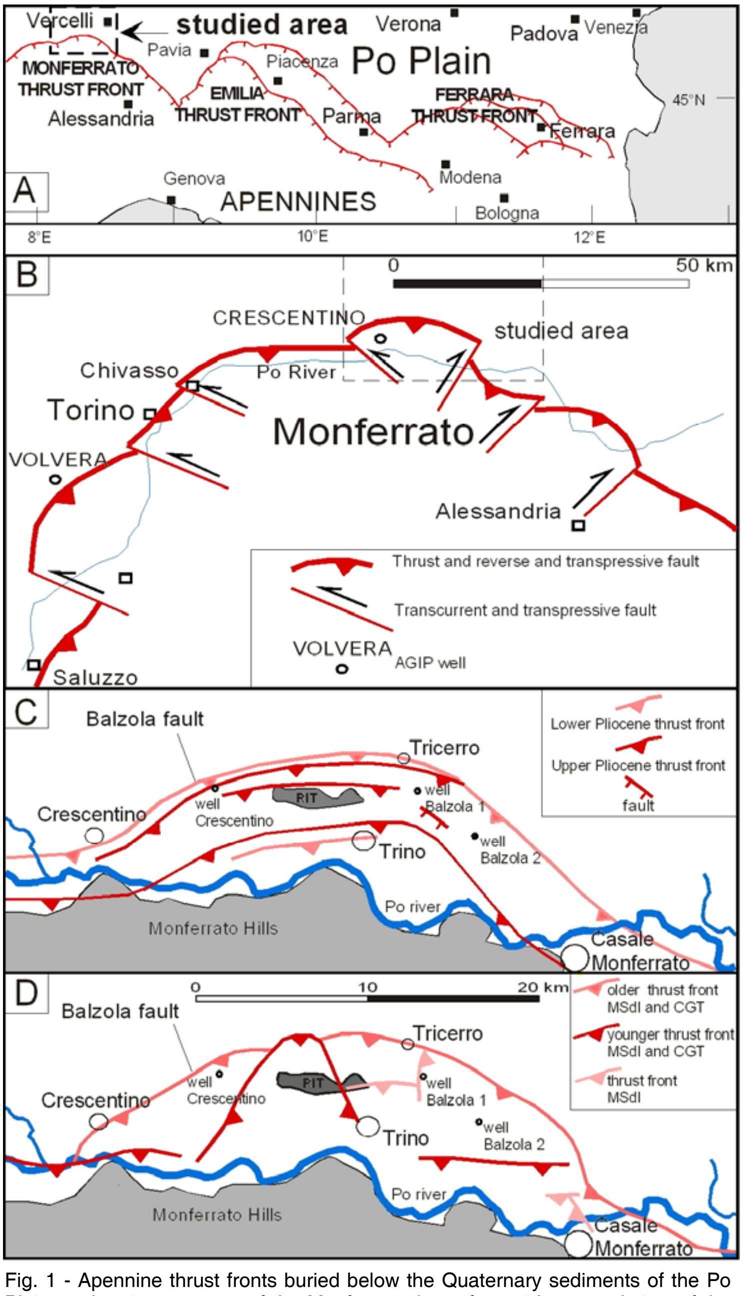

Fig. 1 - Apennine thrust fronts buried below the Quaternary sediments of the Po Plain, and main structures of the Monferrato thrust front. 1A: general view of the Apennine buried fronts. 1B: the Monferrato-Torino Hills thrust front according to Costa (2003). 1C: the Monferrato thrust fronts below the Quaternary sediments of the Vercelli Plain, according to the Geological Map of Italy F. Vercelli (1969) and ENEL (1985). 1D: the Monferrato thrust fronts below the Quaternary sediments of the Vercelli Plain according to (MSdI = Bigi et al., 1990) and (CGT = Dela Pierre et al., 2003a).

3. METHODS

In this paper a detailed discussion of the topographical, geomorphological and geological elements is presented in order to assess the chronology of the tectonic phases and to verify whether the most recent tectonic activity is actually connected with the Monferrato thrust front or with other structures. Topographic observations are based on the trend of the contour lines reported (Ajassa et al., 1990) on the Elevation Map of the Piedmont Plain (Carta altimetrica dell'alta pianura

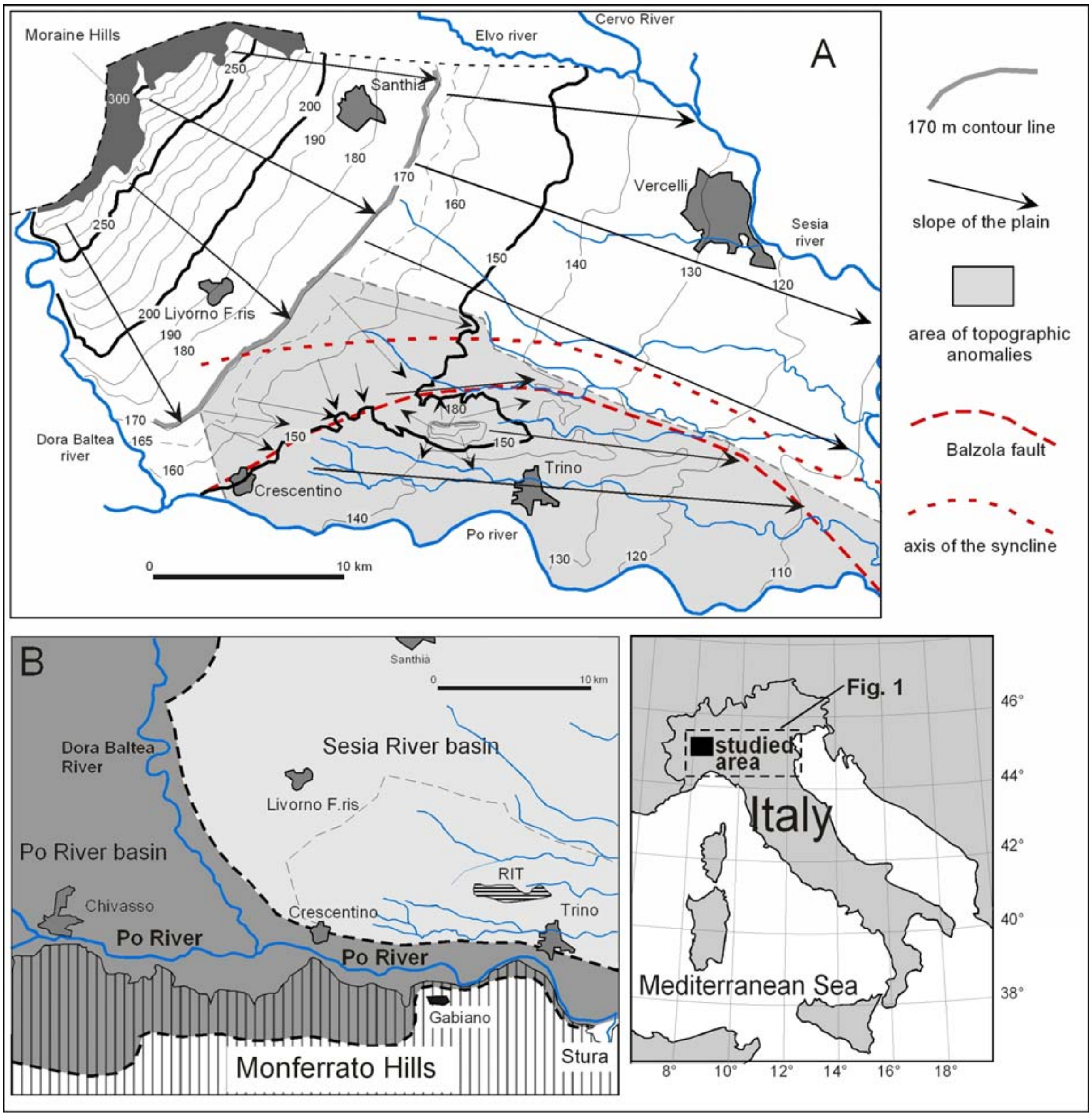

Fig. 2 - Location of the studied area. 2A: contour map, slope of the plain, topographic anomalies and main geological structures buried below Quaternary sediments. 2B: the river Po catchment basin west of the confluence of the Stura stream.

piemontese in Italian) at the scale of 1:250,000. The data used to draw the topographical profiles are reported on the Carta Tecnica della Regione Piemonte (the official regional topographic cartography) at a scale of 1:10,000.

The profiles were designed using 5 m contour lines. The geological maps presented are the result of both field surveys carried out over a long period of time, and of the interpretation of aerial photographs. The flight done in the '50s (Istituto Geografico Militare, flight) was mainly used, as it is the most useful one for the recognition of low terrace scarps. As a matter of fact, agricultural works carried out in the second half of the past century often erased morphological features that, in turn, could be relevant for the interpretation of the evolution of the area.

The lithology of the sediment forming the surface of the terraces has been observed on terrace scarps, along the banks of streams and on ephemeral exposures. Field data have been integrated with stratigraphic data from boreholes. In the area west of Ronsecco and NW of the RIT, the stratigraphy of the sediments was surveyed along several 2.5-m-deep, and 100-m-spaced trenches, all dug for the foundation of high voltage pylons, and arranged along two different alignments, for a total length of 7-8 km.

The chronological framework was mainly based on the morphological correlation between the terraces in the Vercelli Plain and the glacial and glaciofluvial features of the Ivrea moraine amphitheatre (Gianotti et al., 2008). The age of some sediments has been constrained by previous geochemical characterization of a tephra layer (ENEL, 1984), by the dating of a second tephra layer with the fission track method (ENEL, 1984), by some radiocarbon dating (Tropeano & Olive, 1989; Giraudi, 1998) and by the presence of archaeological artefacts from the Lower Palaeolithic to the Bronze Age (Fedele, 1976; GSQP, 1976; Giraudi, 1998).

I have assessed the stratigraphy of the Quaternary deposits through both the stratigraphy of water wells (Varalda et al., 2006; ARPA Piemonte, 2014), and direct observation of boreholes drilled in the framework of a study for the seismotectonic characterization of areas identified as possible locations of nuclear power plants, in the years between 1976 and 1984 (ENEL, 1984).

The ENEL data have already been used for the geological map of Italy at a scale of 1:50,000 (Dela Pierre et al., 2003a) and the stratigraphies of the boreholes carried out by ENEL are listed and described in the Geotechnical Data Bank of the ARPA Piemonte (2014).

In order to infer the Quaternary tectonic deformations, thickness changes of alluvial and glaciofluvial sedimentary units and different elevations of the base of sediments forming the terraces have been evaluated.

Indeed, the stratigraphic data show a great thickness variability of glaciofluvial and alluvial sediments forming the plain North and South of the Monferrato buried thrust front.

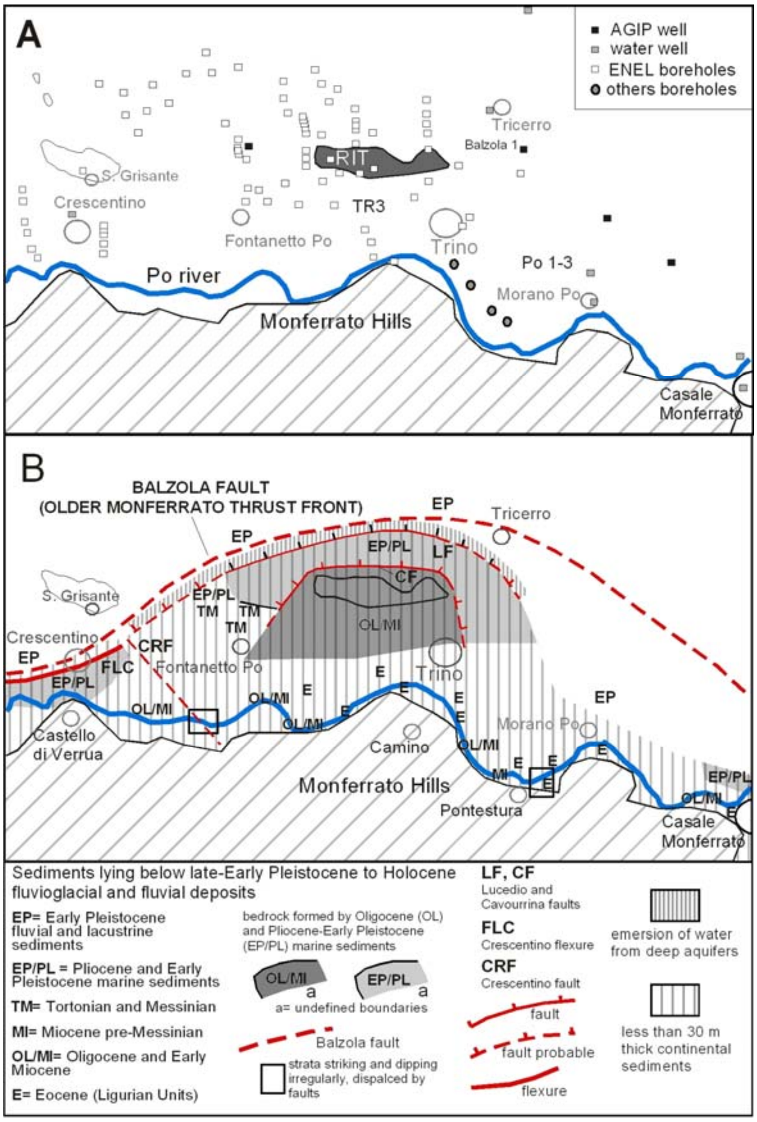

Fig. 3 - Position of the boreholes and the wells, Tertiary and Early Pleistocene sediments lying below late Early Pleistocene to Holocene deposits. 3A: boreholes and wells used for their stratigraphic data. 3B: age of the sediments sampled in the boreholes covered by glaciofluvial and fluvial deposits, and outcropping in the Po river bed, main tectonic structures inferred from the borehole stratigraphy.

4. GEOLOGY OF THE VERCELLI PLAIN

The Vercelli plain is composed, almost completely, by glaciofluvial, fluvial and aeolian Quaternary sediments (SGd'I, 1969a). Marine tertiary sediments outcrop on the scarps of the highest terraces and in the bed of the river Po, i.e. where the river is eroding the northern slope of the Monferrato Hills.

As already mentioned, the southernmost portion of the plain is formed by Quaternary sediments, less than 15-20 m thick, lying on the Tertiary bedrock deformed by folds, fault and involved by the frontal thrust.

The following description of Tertiary sediments and tectonic structures is based on data derived from geophysical researches and wells dug during oil exploration campaigns (Pieri & Groppi, 1981; Cassano et al., 1986) and the seismotectonic characterization of sites suitable for a nuclear power plant (ENEL, 1984). Lastly, data from the Po river bed derive from a survey of the outcrops.

4.1. Tertiary bedrock and main tectonic structures

Previous works on the study area have stated the bedrock of the southern Vercelli plain is formed by Tertiary marine sediments involved in one or more thrust fronts. Tectonic interpretations (Fig. 1C; 1D) reported in SGd'I (1969a), ENEL (1984) and in Bigi et al. (1990) show that the northernmost thrust front (Balzola fault) affects at depth the Lower Pliocene marine sediments. South of the Balzola Fault, one or more thrust fronts are reported by the Authors. According to Costa (2003), the northernmost part of the Monferrato thrust front has been extruded northwards along two lateral structures (Fig. 1B) located east and west of the front.

Borehole data (ENEL, 1984) aided in recognizing the presence of marine sediments below the Quaternary glaciofluvial and fluvial sediments, mainly dated from Eocene to Pliocene, but also to the onset of the Early Pleistocene, and of Early Pleistocene alluvial and lacustrine sediments (Fig. 3A, B). In this paper the Upper Pliocene marine sediments reported in ENEL (1984) are dated at the Early Pleistocene of the new international geochronological scale (Gibbard et al., 2010).

Except for some areas (Oligocene, Pliocene and Early Pleistocene marine sediments outcropping or drilled in the RIT area and south of Crescentino) the boundaries between the different marine formations cannot be clearly traced. It follows that in Fig. 3 the sediments and their ages are shown with symbols in correspondence to the places where they have been recognized.

Through borehole alignments (ENEL, 1984) it was possible to deduce the presence of at least two faults (Cavourrina and Lucedio Faults - Fig. 3A), the position of which lies within a 40-50-m-width strip matching the distance between the closest pair of boreholes.

The Cavourrina Fault juxtaposes Oligocene and Pliocene marine sediments (Fig. 3B). The fault bounds the northern edge of the RIT and it was very likely responsible for its uplift. West of the RIT, the fault plane dips to the NW. North of the RIT, according to the drilling stratigraphy, the fault plane is nearly vertical. Southeast of the RIT, the fault reaches at least the latitude of Trino (Fig. 3), where two closely spaced drillings show the presence of marls, probably Oligocene (in the west) and of Pliocene silty sands (in the east). There are no data that can be used to infer whether the fault continues to the south, outside of the area surveyed with boreholes.

The Lucedio Fault trends WSW-ENE, being slightly convex northward. Here Pliocene and Lower Pleistocene marine sediments are faulted against Lower Pleistocene fluvial and lacustrine sediments, inducing the subsidence of the northern area at least after the Lower Pleistocene. In the alignment where the boreholes are closer, the northward dip of the fault plane cannot be less than 80°.

Hydrogeological studies in the area (ENEL 1984) have shown that, just to the north of the Lucedio Fault, there are springs fed by water rising from deep aquifers (Fig. 3) having a chemical and isotopic composition very different from that of water of surrounding areas.

The rise of deep water also occurs NW and SE of the stretch of the fault indicated by an alignment of boreholes: it is therefore probable, as suggested in ENEL (1984), that the Lucedio fault might reach the areas north-east of Crescentino and of Salera, south-east of Tricerro.

A seismic line crossing the faults north of the RIT (ENEL, 1984) shows that both the Cavourrina and Lucedio Faults match the thrust fronts evidenced by the SGd'I (1969a) (Fig. 1). Here the Cavourrina Fault could partly fit in with the youngest thrust front reported by Bigi et al. (1990).

On the other hand, ESE and WSW of Crescentino, where many authors placed the Monferrato thrust front (SGd'I, 1969a; Pieri & Groppi, 1981; ENEL, 1984; Cassano et al., 1985; Dela Pierre et al., 2003), at least down to the depth of about 200 m, any of the available borehole logs account for the existence of faults similar to the Lucedio one. The Pliocene and Lower Pleistocene marine sediments dip 10-15° to the NNW and are deformed by a flexure. The flexure starts just east of Crescentino near the western termination of the Lucedio Fault.

The change of the tectonic features of the thrust front east of Crescentino implies the presence of a fault crossing the front (Crescentino Fault) which could match the NW-SE lateral ramp hypothesized by Costa (2003). Indeed, the Crescentino Fault (Fig. 3B) reaches the Po river bed in a place where outcropping Tertiary marine sediments show NW-SE fault planes.

In the area between Pontestura and Morano Po, the river bed is almost completely carved in Tertiary bedrock (Fig. 3B): the presence of calcareous marl suggests that it belongs to the Eocene terms of the "Ligurian" units, the same reported by Dela Pierre et al. (2003a) in the Po bed south of Trino. In the river bed NE of Pontestura, nearly vertical marl strata, running N-S, outcrop, while to the east strata dip very irregularly and are displaced by faults. Undeformed NNW-SSE marl outcrop in the river bed SW of Morano Po, that is, east of the area where the sediments are displaced by faults.

Also near Casale Monferrato, in the river bed, Eocene clay and calcareous marl of the "Ligurian" units outcrop, although a few dozen metres north of the river bed, Pliocene marine sediments (subzone MP14 and MP15a) have been found in boreholes (Violanti & Sassone, 2008). The boundary between Eocene and Pliocene sediments is not exposed, but strongly deformed outcropping sediments suggest the presence of an unconformity or a tectonic contact.

The geological structure of the Casale Monferrato area matches that of the Castel Verrua area, South of Crescentino, where Pliocene sediments unconformably overlie the deposits of the "Complesso Indifferenziato" ("Ligurian" units; Zappi, 1961; Dela Pierre et al., 2003a,b). At Casale Monferrato, the contact could correspond to the projection at the surface (Fig. 1C) of a NW-SE thrust front (evidenced in ENEL, 1984) or of another front (Fig. 1D) hypothesized in Bigi et al. (1990).

4.2. Geomorphology and Quaternary stratigraphy of the Vercelli Plain

The Vercelli Plain is quite homogeneous, especially in the north-western and central-northern areas (Fig. 2A). According to all previous works (SGd'I, 1966; SGd'I, 1969a; GSQP, 1976; Carraro et al., 1975; Giraudi, 1998; Dela Pierre et al., 2003a; Gianotti et al. 2008) it is mainly formed by glaciofluvial and fluvial sediments dated between the Mindel glaciation and the Holocene.

Despite the scarcity of exposures, the stratigraphy of the sediments is fairly well known thanks to boreholes drilled for water and stratigraphic surveys. The stratigraphic data enable the thickness of the Quaternary sediments to be evaluated and also to differentiate the areas in which sedimentation prevailed from others where erosion and sedimentation phases occurred.

Based on the abundance of terraces formed by the main streams, (Fig. 4) the Vercelli Plain can be divided

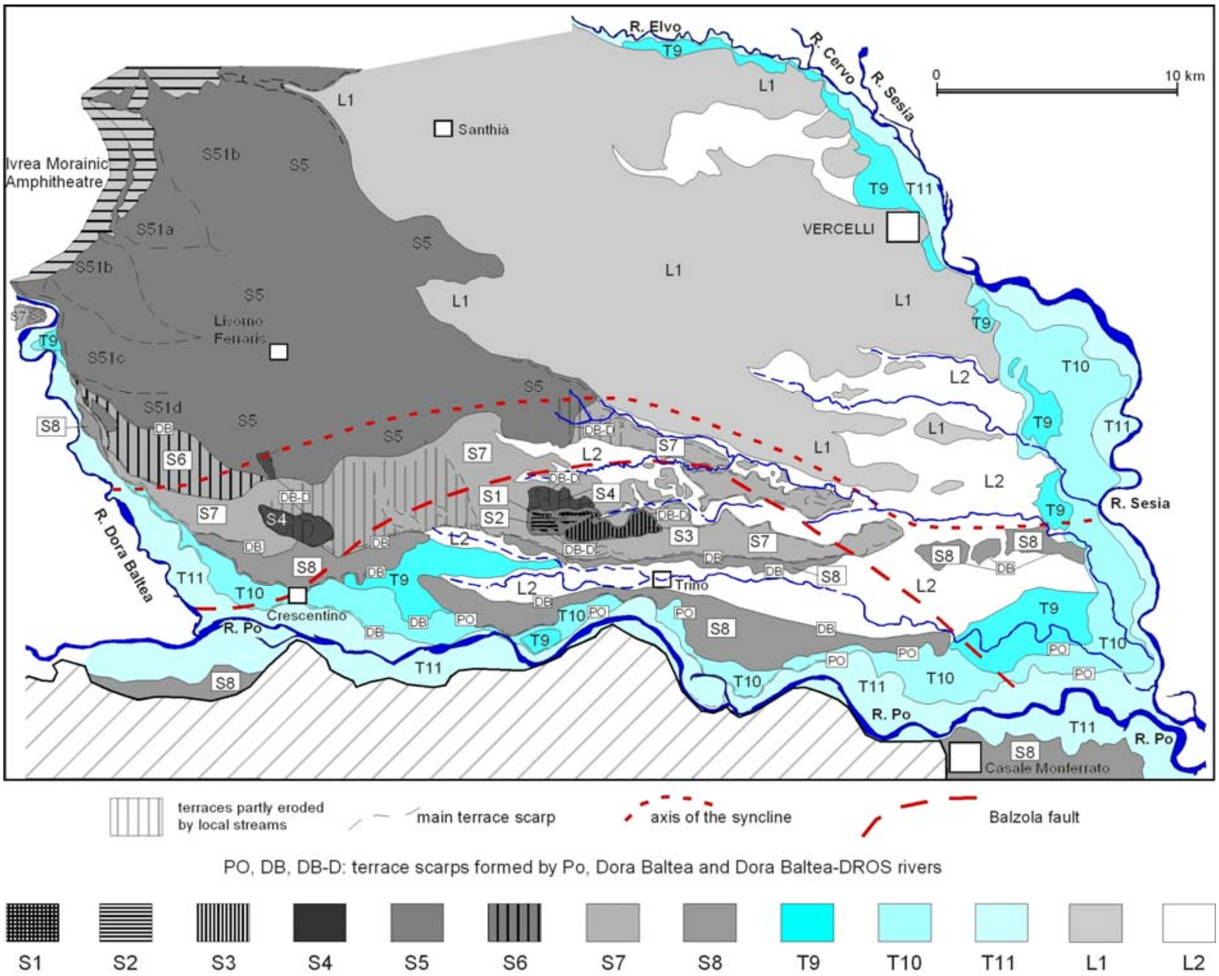

Fig. 4 - The terraces of the Vercelli Plain. S1-S3: terraces forming the Trino isolated hill (RIT). S4-S8: terraces shaped by Alpine rivers when their bed was far from the modern ones. T9-T11: terraces forming a belt near the modern day Alpine river beds. L1-L2: terraces shaped by local streams.

into four parts: the north-western area, formed by some terraced surfaces, is a narrow strip surrounding the hills of the SE Ivrea morainic amphitheatre. The central-northern area, extending over about two-thirds of the studied plain, forms the main surface of the plain. The southern area is formed by seven different terraces while the last well recognizable area is the belt formed by three terraces parallel to the river beds.

The four older terraces (S1 to S4) are only in the southern area, while the others (S5 to S8 and T9 to T11) are more diffuse. The stratigraphy of the sediments forming the terraces will be described from the oldest to the most recent. However, some parts of the Vercelli Plain are formed by terraces shaped by local streams (L1-2), and will be described separately.

The relationship between sediments that form terraces and moraines of the Ivrea amphitheatre indicates the glaciofluvial origin of sediments and suggests the age of some terraces.

The bottom of sediments forming the oldest terrace is carved into the Tertiary bedrock, while the bottom of other glaciofluvial deposits cuts into older sediments and sometimes into the Tertiary bedrock. As determined in the whole of the Mediterranean basin by Macklin et al. (2002), I have assumed that, if the phases of coarse sedimentation predominantly correspond to periods of glacial expansion, the phases of prevailing erosion mainly correspond to interstadials or interglacials. The terraces (Fig. 4), due to their characteristics, can be divided into groups. One group (S1-S3) includes only the terraces that form the RIT, relating to a landscape that was significantly different to the modern one. Conversely, the distribution of younger terraces (S4-T11), the inclination of their surfaces, and the direction of their scarps enable the identification of their own stream. In this group of terraces a first subgroup (S4-S8) can be distinguished: this includes terraces shaped by rivers that flowed, at least in certain parts, in areas away from the current river beds. The terraces form the eastern portion of the river Dora Baltea glaciofluvial fan, and the area, oriented approximately W-E, between the RIT and the Monferrato hills. A second subgroup includes terraces (T9-T11) lying along the modern river beds.

Several exposures and the stratigraphy and lithology

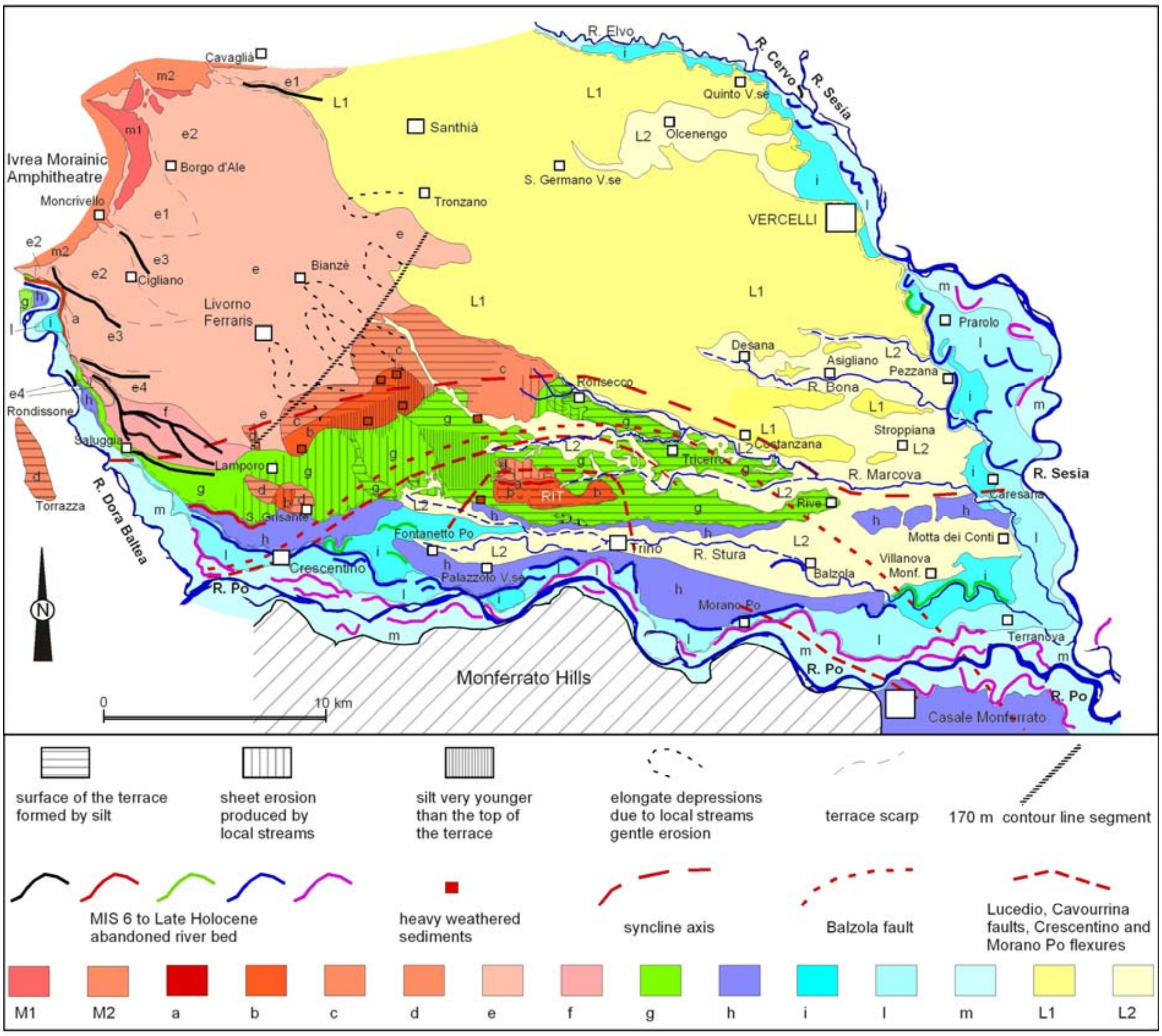

Fig. 5 - Geological map of the Vercelli Plain. Legend- m1-m2: moraine ridges of the Serra Alloformation; a-h: late Early Pleistocene to late Upper Pleistocene glaciofluvial sediments; a: late Early Pleistocene (MIS 22-?); b: late Early Pleistocene to Middle Pleistocene (MIS 22?-MIS 12); c: Middle Pleistocene (MIS 10); d: Middle Pleistocene (MIS 8); e: late Middle Pleistocene (MIS 6); f: late Middle Pleistocene or early Upper Pleistocene; g: early (?) and middle Upper Pleistocene (MIS 4?); h: late Upper Pleistocene (MIS 2); i, l, m: Late Holocene. L1: sediments from Valdora temporary stream and other local streams (MIS 4? - MIS 2?); L2: sediments of the local streams (MIS 2?-Late Holocene).

of sediments obtained from boreholes show that some terrace surfaces are formed by sediments of different ages.

Terrace S1 - Sediments that form terrace S1 ("a" in Fig. 5) are described in SGd'I (1969a), GSQP (1976) and ENEL (1984) as directly overlying the Oligocene marine bedrock, 8-9 m thick, and formed, at the bottom, by glaciofluvial sandy gravel. Sediments are very weathered (colour 2.5-5 YR of the Munsell Soil Color Chart, MSCC). The degree of weathering of sediments is also similar to that of deposits partly outcropping on a terrace scarp eroded by the Dora Baltea river ("a" in Fig. 5), not far from the morainic front, and assumed to be formed before the Mindel glaciation (SGd'I., 1969a). A layer of pedogenized loess (coloured 5YR, MSCC) lays above sandy gravel, and it is covered by a younger, pedogenized loess (coloured 7.5YR, MSCC). According to the SGd'I (1969a), the older loess sedimented during the Mindel glaciation, while the second one is dated at the Riss glaciation. Another thin and discontinuous loess horizon, (pedogenized and coloured 10YR, MSCC), ends the stratigraphic sequence.

The S1 glaciofluvial sediments are weathered and pedogenized to the same degree as some morainic deposits in the Northern Ivrea amphitheatre reported by Gianotti et al. (2008). According to Carraro et al. (1991) the moraines with soils coloured 2.5YR (MSCC) are the oldest and, according to their palaeomagnetic characteristics, can be dated to the late Early Pleistocene. Other moraines, early Middle Pleistocene in age (Gianotti et al., 2008), contain soils coloured 5YR (MSCC). The S1 glaciofluvial sediments, being the oldest and most weathered, could have the same age as the oldest moraines and date back to late Early Pleistocene.

The loess dating is based mainly on the presence of prehistoric artefacts. The younger loess covers some Upper Palaeolithic artefacts and can be dated at the MIS 2. The intermediate loess covers Lower Palaeolithic artefacts and includes at the top some Middle Palaeolithic Mousterian artefacts. Thus, the top of the loess may be dated to the late Middle Pleistocene or Upper Pleistocene (Fedele, 1976), that is, at MIS 6 - MIS 3. The older loess is, therefore, older than MIS 6 and dated at least at MIS 8.

Terrace S2 - The terrace surface S2 is separated from that of S1 by a scarp that reaches a height of approximately 10 m. The stratigraphy of the sediments forming the terrace (SGd'I, 1969a; GSQP, 1976; ENEL, 1984), is very similar to that of S1. Glaciofluvial sandy gravel and very weathered sand (coloured 2.5-5YR, MSCC), ("b" in Fig. 5) form the main sedimentary body. The S2 sediments, assumed to be older than the Mindel glaciation by SGd'I (1969a), can be correlated to the moraines having the same degree of weathering and can be dated to the late Early Pleistocene or the early Middle Pleistocene. They are covered by two loess horizons, the older and more weathered one (colour 5YR) being covered by a less weathered one (colour 7.5YR). At the top of the younger loess, Middle Palaeolithic Mousterian artifacts have been found (Fedele, 1976). We can assume, therefore, that the loess horizons correspond to those described for the S1 surface and that they are of the same age (MIS 6- MIS 3 and MIS 8). On the whole, the sediments are 15-16 m thick.

Terrace S3 - Terrace surface S3 is separated from that of S2 through a scarp that reaches a height of approximately 10 m. The stratigraphy sediments (GSQP, 1976; ENEL, 1984), is made by gravelly sand and sandy gravel, with interbedded sand and silty sand layers, weathered at their top and coloured 5YR (MSCC). Also these sediments are originally glaciofluvial, but interbedded fine sediments are colluvia that originated from the S2 terrace scarp. The glaciofluvial sediments can be correlated to moraines having the same degree of weathering and, consequently, dated to the early Middle Pleistocene. As on terrace S2, the pedogenized glaciofluvial sediments are capped by two loess horizons: the older one is more weathered (5YR in colour) than the younger one (7.5YR). The loess horizons can be correlated, therefore, to those lying on terraces S1 and S2. On the whole, the thickness of sediments forming the terrace S3 can be around 15-17 m.

Terrace S4 - The terrace S4 has been observed in three places, north of the RIT and north of Crescentino. Here the S4 surface is preserved in two small, isolated flattened heights which are surrounded by younger terraces. The first, larger one (nearly divided into two parts by a small valley with a flat bottom) lies near San Grisante, while the second one is near Lamporo. Sediments forming the San Grisante terrace (GSQP, 1976; ENEL, 1984), are glaciofluvial sandy gravel covered by silt and sandy silt. Glaciofluvial sediments ("b" in Fig. 5) are very weathered and there is a truncated palaeosoil at their top. Therefore, between the end of the glaciofluvial sedimentation and the onset of the deposition of top silts, a strong pedogenetic phase occurred which was followed by an erosion. The degree of weathering of San Grisante glaciofluvial sediments is similar to that of sediments forming terraces S2 and S3 and of moraines dating back, according to Carraro et al. (1991) and Gianotti et al. (2008), to the late Early Pleistocene or the early Middle Pleistocene. According to GSQP (1976) the soil on the top of the glaciofluvial sediments is the so-called "ferretto", which developed during the interglacial between the Mindel and the Riss glaciations, that is during the MIS 11. The glaciofluvial sandy gravel should therefore have been sedimented during the MIS 12 or earlier.

Silt and sandy silt sediments, overlying the top of glaciofluvial sandy gravel, form a layer of variable thickness (from 3.6 to 6 m) and are capped by a soil coloured 10YR/7.5YR ("d" in Fig. 5). Silt is similar to that forming the top of the Torrazza-Rondissone terrace, lying W of the Dora Baltea river, described by GSQP (1976). Silt sedimented on a flood plain or in a palustrine basin, but at its top it consists of loess-derived deposits. In the Torrazza-Rondissone terrace, interbedded between the silty deposits, some 0.1-0.5-m-sized polished and striated pebbles form some stone-lines. According to GSQP (1976) pebbles could have been carried there by floating ice chunks stemming from the tongue of the Dora Baltea glacier. The sin-glacial origin of sediments assumed by SGd'I (1969a) is therefore confirmed. In a quarry of the Torrazza-Rondissone terrace, interbedded in the upper silt sediments, a horizon containing volcanic glass was sampled. Mineralogical characteristics of glass suggest that it originated from the M. Amiata volcano (ENEL, 1984), which was active between 300 and 180 ka BP (Bigazzi et al., 1981), that is, in a period including the MIS 8 and the early MIS 6.

North of the RIT, top sediments forming the surface are made by 4-to-6-m thick silt, very similar in its lithology and sedimentary facies to that described above and containing remnants of the tephra layer. Silt covers unweathered sandy gravel ("d" in Fig. 5), about 8 m thick. Below the sandy gravel, a colluvium of a soil formed by very weathered silty clay coloured 2.5YR, was sampled in a borehole. It follows that the colluvium derives from the top of the pedogenized sandy gravel forming S1, and that the sandy gravel was probably sedimented during the same cold stage just before the silt. As the top of terrace S4 is higher than the S5 (dated at the MIS 6; see below), it was formed in a cold stage preceding the MIS 6. Based on the volcanic glass chronological framework, the cold stage is, therefore, the MIS 8 (280-240 ka BP). The loess-derived palustrine sediments could therefore be correlated with the oldest loess, which overlies soils coloured 2.5-5YR, identified on terraces S1, S2 and S3.

Terrace S5 - This forms the NW part of the Vercelli Plain and derives from the melting of remnants of different terrace surfaces that can be distinguished only near the moraine hills.

The terraces consist of glaciofluvial sediments weathered by a soil coloured 7.5YR. They outcrop discontinuously and, generally, are limited in extent, forming an apron around the moraines. In turn, the moraines lying at the NW boundary of the study area are formed by two differently preserved series of ridges (m1 and m2 in Fig. 5). The presence of a series of terraces and moraine ridges shows that the evolution of the glacial front in this area was very complex.

According to Gianotti et al. (2008), both moraines and glaciofluvial sediments, dated to the Riss glaciation in former geological maps and papers (SGd'I, 1966; 1969a; Carraro et al., 1975), pertain to the "Serra Alloformation" dated to the late Middle Pleistocene, that is, the MIS 6.

To the south and SE the different surfaces join to form a single surface (S5, Fig. 4).

Above the contour line 170 m a.s.l. (Fig. 5), surface S5 is formed mainly by sandy gravel sediments, sometimes covered by a thin silt layer, weathered by a soil coloured 7.5YR. At a lower altitude, surface S5 develops on sediments having different ages and origins.

In the area NE of Lamporo, glaciofluvial sandy gravel, heavily weathered (coloured 2.5-5YR) down to a depth of 10-12 m ("b" in Fig. 5), is both exposed and sampled in some boreholes; this is very similar to that underlying the silt that forms the S4 surface near San Grisante. The degree of weathering of sediments is similar to that of S2 and S3 terraces and of the moraines dating back to the late Early Pleistocene or the early Middle Pleistocene (Carraro et al., 1991; Gianotti et al., 2008). In the area NE of Lamporo, pedogenesis lasted for a longer period than in the San Grisante and RIT area, and has continued till the present day. Arduino et al. (1984), studying the relationships between the age of soils, the presence of iron oxides and the redness of soils, evidenced that the pedogenesis did not stop during cold periods.

Other boreholes showed that at the northern and eastern boundaries of the area with strongly weathered sandy gravel, there are sandy gravelly sediments a few metres thick, less weathered and coloured 7.5YR ("c" in Fig. 5). These sediments are covered by the younger glaciofluvial sediments that were described above, dated to the MIS 6.

In some places surface S5 is capped by discontinuous layers of silt, 2-3 m thick, alluvial in origin, formed by the sedimentation of material derived from the erosion of soils developed on glaciofluvial deposits. The silt can be both stiff and weathered by a soil coloured 7.5YR, or less stiff, but less pedogenized (soil coloured 10YR). The latter sometimes overlaps the more weathered one. In profiles 1 and 2 (Fig. 6) it can be noted that the boundaries between the different sedimentary units described above are evidenced by small changes in the slope of the plain: the places where the most weathered sediments are exposed are the flattest.

Terrace S6 - This terrace is only in the western plain (Fig. 4). A scarp 3-4 m high separates it from S5 but it becomes less clear and nearly disappears towards the SE. The terrace is made by sandy gravel ("f" in Fig. 5), the thickness of which is unknown, younger than the MIS 6 glaciofluvial sediments forming S5.

Terrace S7 - The surface of terrace S7, surrounding both the RIT and the S4 terraces, is well represented in the area. The scarp that separates the terrace from S6 is higher to the west and becomes smaller towards the east, and the boundary between surfaces S7 and S5 is not a real scarp, but only a slight break in the slope of the plain (profiles E, F, in Fig. 6).

West of the Lamporo area, sediments forming the terrace ("g" in Fig. 5) are mainly glaciofluvial sandy gravel, while to the East, in places where the surface is less eroded by local streams, a layer of alluvial silt, 2-3 m thick, pedogenized by a soil coloured 7.5YR, covers the sandy gravel. In a few places, the top silt is capped by another layer of silt not so stiff, with a less evoluted soil coloured 10YR, which can be correlated to the younger silt layer lying on terrace S5. The sediment thickness is on the whole about 15 m.

Near the SW slope of the RIT, below the silt that forms surface S7, sandy gravel weathered by a soil coloured 5YR has been found. These sediments are probably remnants of the sandy gravel forming also terrace S3, only eroded at the top during the shaping of terrace S7.

The terrace correlates with that evidenced by Gianotti et al. (2008), outside the moraine amphitheatre, on the right of the Dora Baltea river, and formed by glaciofluvial deposits of the Piverone Alloformation. The terrace sediments, therefore, can be dated to the period following the MIS 6 glacial maximum (which ended about 140 ka BP) and preceding the MIS 2 glacial maximum (which started about 30 ka BP), named MIS 4? in this paper.

Surface S7 was eroded by local streams. West of the RIT, these streams caused mainly the erosion of the summit silt, while north and NE of the RIT, the terrace is cut by a complex network of small valleys, the bottom of which lies between 1-2 m and 7-8 m below the terrace surface.

Terrace S8 - The scarp dividing S7 from S8 is lower towards the west and higher towards the east: it is this feature that makes it different from those that separate the terraces S5 from S6, and S6 from S7. In the plain south of the Po river, terrace S8 has been observed only west of Castel Verrua and east of Casale Monferrato, but is lacking in the intermediate area.

Sediments forming the terrace ("h" in Fig. 5) mainly consist of sandy gravel, overlain by a soil coloured 10YR. The sediment thickness, which is known only in the Trino and Palazzolo area, does not exceed 10 m.

The terrace correlates with that evidenced by Gianotti et al. (2008), outside the moraine amphitheatre, on the right of the river Dora Baltea, and is formed by glaciofluvial deposits of the Ivrea Alloformation. Inside the amphitheatre sediments of the Ivrea Alloformation are dated to the last glacial maximum, around 21 ka BP, according to Gianotti et al. (2008), or are younger, because they also lie upstream of the terminal moraines. The sediments forming S8 can be dated, therefore, to the late Upper Pleistocene, between 21 and 14-15 ka BP.

Terraces T9, T10, T11 - These are the most recent terraces and form a discontinuous strip along the river beds. It should be noted that starting from the area south of Fontanetto Po down to Trino, the strip formed by terraces T9, T10 and T11 is narrower than in the western and eastern areas (1.2 to 2.5 km wide, against the 2.5-5 km near Crescentino and 3-3.5 km east of Morano Po). In the plain south of the Po, between Verrua Savoia and Casale Monferrato, only terrace T11 is present.

Alluvial sediments forming terraces T9, T10 and T11 are mainly of sandy gravel lithology (i, l, m in Fig. 5). Their thickness, visible only between Fontanetto Po and Trino, does not exceed 6-8 m. In the sediments that form terrace T9 some useful dating elements have been found. In a borehole drilled near Fontanetto Po an iron nail contained in the sandy gravel came to light, similar to those used by Romans for the Ivrea harbour in the Dora Baltea river (F. Gambari, personal communication). In a quarry South of Villanova Monferrato, tree trunk remnants have been exposed, dated 380-600 AD, 115-725 AD and 440-640 AD (2σ calibrated ^{14}C age; Tropeano & Olive, 1989). It follows that terrace T9 should be dated to the Roman and the Early Medieval periods.

On the Po banks, near Palazzolo, Trino and Morano Po, where sandy gravel forming terraces T10 and T11 lie directly on the Tertiary bedrock, at the bottom of the alluvial sediments, bronze axes and swords from the Bronze Age have been found (Janigro D'Aquino,1979; Fozzati & Giraudi, 1983; Facchin, 1997; Giraudi, 1998). Near the Po bed, north of Camino, the T11 sediments cover the remains of a medieval building and contain remnants of poles, from the Middle Ages (1225-1400 and 1325 -1460 AD), and trees, dated to the Renaissance (1410-1620 AD) (2σ calibrated ^{14}C age; Giraudi, 1998). On the whole we can state that sediments forming the three most recent terraces date back to the last 3-4 millennia.

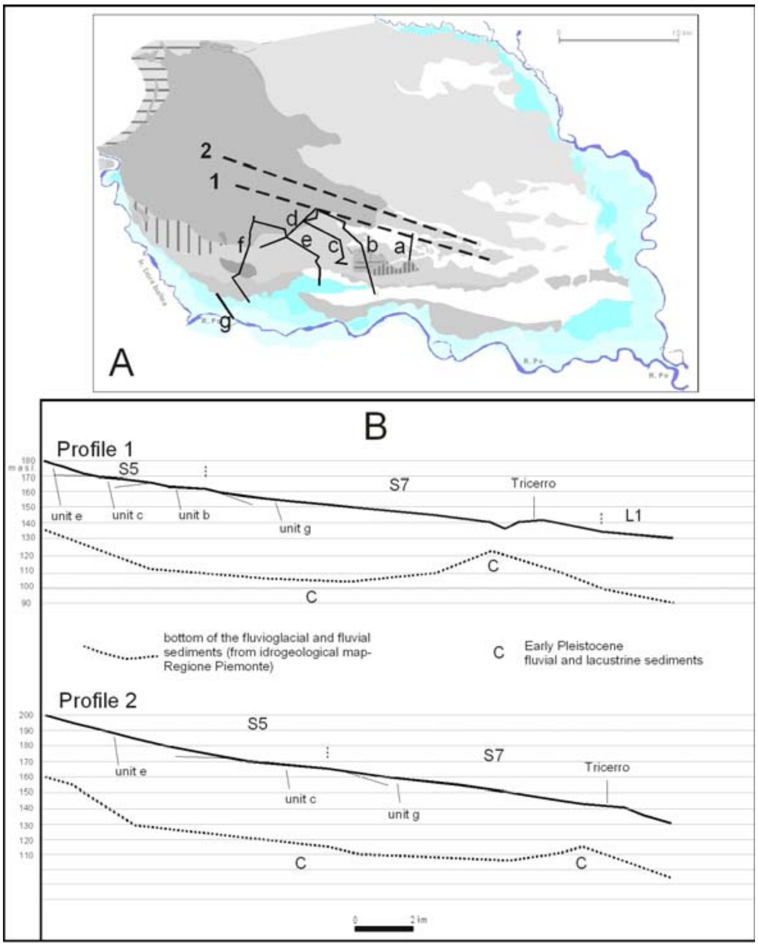

Fig. 6 - Profiles along the Vercelli Plain. 6A: terrace map with the position of profiles; 1, 2: profiles reported on Fig. 6B; a,b,c,d,e,f: position of the profiles reported in Fig. 7; 6B: profiles showing the slope of the plain, the boundaries between terraces and stratigraphic units and thickness of glaciofluvial sediments lying on Early Pleistocene fluvial and lacustrine deposits.

Terraces shaped by small local streams - The Vercelli plain has been shaped also by local streams mainly fed by water from springs sited along the "risorgive" area (De Luca et al., 2005). Fig. 4 shows two areas shaped in a different way by said streams.

Surface L1 - This is the surface, lower than S5 and S7, that forms the central and Northern Vercelli Plain, formed mainly by sandy gravel. L1 is fairly regular but in the eastern part it appears cut by small valleys with flat bottoms, 1 to 3 m deep, produced by smaller local streams.

The scarp between S5 and L1, well recognizable in the NW area, becomes difficult to identify towards the SE, but NW of Ronsecco it is again well recognizable. In the north-western plain, L1 narrows and enters an abandoned river bed, produced by a glacial meltwater stream, which cut into the MIS 6 glaciofluvial sediments. The stream that shaped the L1 surface, therefore, came from the former stream bed (Fig. 4), called Valdora. According to Gianotti et al. (2008) the glacial meltwater stream was linked to the Piverone Alloformation glacial phase, doubtfully attributed to MIS 4, and was therefore contemporaneous with the glaciofluvial sedimentation forming the S7 terrace. However, the surveys carried out for this paper have evidenced that the bottom of the former Valdora bed, formed by sandy gravel having a homogeneous grain size and lithology, is cut across by a gently sloping scarp a few metres high. The surface L1 begins east of the scarp, therefore it is younger than the Piverone Alloformation. In the southern Vercelli plain, surface L1 cuts surface S7, which is correlated with said Alloformation, confirming that L1 was younger. The scarp, which crosses the former Valdora bed, cannot have been produced by lithological discontinuities or tectonic displacements, because these would also have affected the surrounding areas formed by older sediments. It seems reasonable to believe the bottom of the former bed was partly modified by the water of a large spring, which formed near the base of the gently sloping scarp, and was active for a period following the

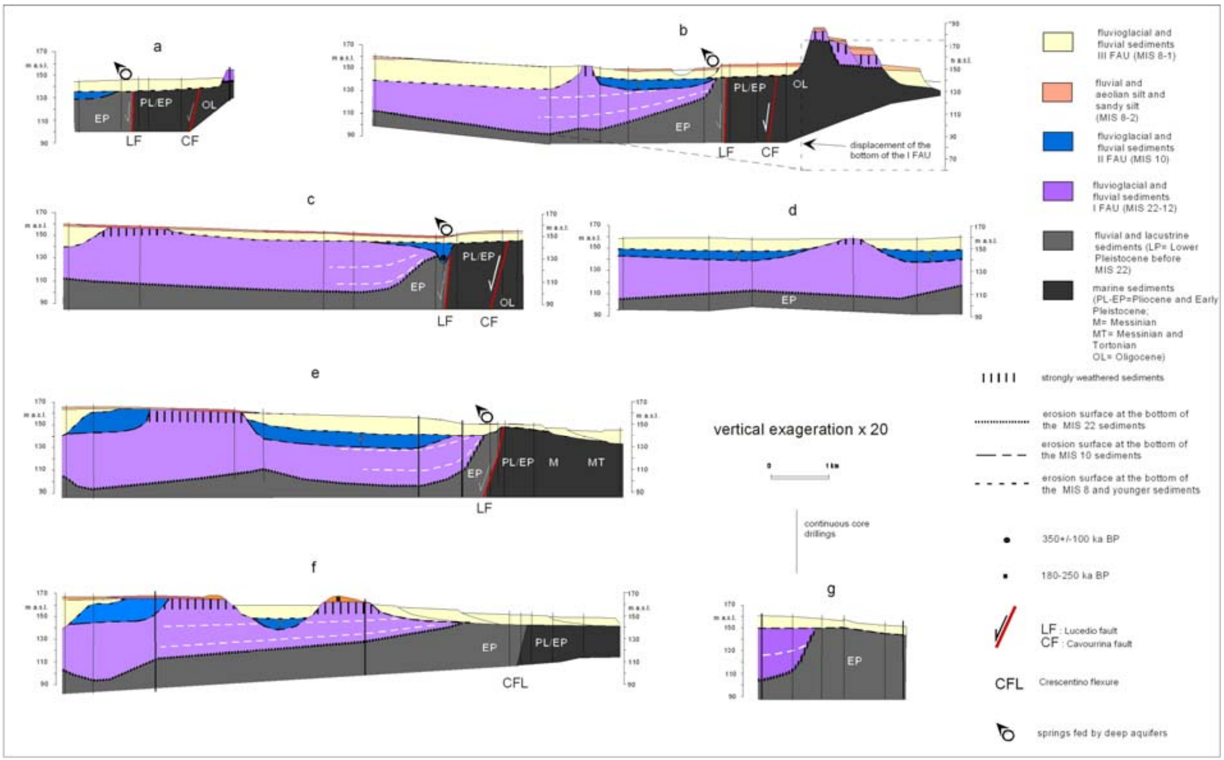

Fig. 7 - Stratigraphic sections of the sediments forming the Vercelli Plain obtained from ENEL (1984) boreholes.

Piverone Alloformation glacial maximum.

Surface L2 - L2 represents a cluster of surfaces shaped by local streams. In fact, the different surfaces are difficult to map because the very low scarps separating them are very difficult to follow and have been partly erased by agricultural works. L2 forms also the bottoms of the small valleys cutting into fluvial terraces and L1 surfaces. The sediments forming L2 are generally silty sand, sand and sandy gravel; their thickness, that reaches 2.5 m, is known only in the Trino area because some trenches some hundreds of metres in length have been made for a channel.

4.3 - Thickness of glaciofluvial sediments according to borehole data

The stratigraphy of sediments lying well below the terrace surfaces in the area North of the Lucedio and Cavourrina Faults and the Crescentino Flexure has been evidenced by boreholes that reached a maximum depth of 200 m (Fig. 7). Based on the borehole alignments, some geological sections have been drawn, revealing the thickness and the relationships between both the glaciofluvial sediments and the underlying older deposits, as also the presence of possible tectonic structures.

Three different glaciofluvial and fluvial units (FAU) have been identified, mainly formed by sandy gravel overlying, through a sharp contact, thinner fluvial and lacustrine sediments or the older FAU deposits. The fluvial and lacustrine sediments are dated at the Lower Pleistocene, and are mainly made by sand, silt and clay, with vegetation remnants and small pieces of wood interbedded with thin sandy gravel and peat layers (ENEL, 1984; ARPA Piemonte, 2014; Varalda et al., 2006).

The 1st FAU includes units “a” and “b” which form the terraces, and overlies Lower Pleistocene sediments through an erosion surface and has a maximum thickness of about 70 m. The top of the 1st FAU is in general eroded and covered by later sediments. When the top of the 1st FAU reaches the surface (on terrace S5) or is exposed in quarries (below the silts forming terrace S4 near San Grisante) sediments are heavily weathered (soils coloured 2.5YR or 5YR). As discussed above regarding terraces S4 and S5, pedogenesis occurred probably during MIS 11 and the top of the glaciofluvial sediments is dated to the MIS 12 or is older.

Based on the age and the weathering, we assume that the 1st FAU can be correlated to the strongly weathered glaciofluvial sediments of the RIT. However, while on the RIT sediments form three different terraces and sedimentary units with a bottom gradually at lower altitudes, north of the Lucedio Fault the 1st FAU formed during conditions of prevailing sedimentation.

There are no direct data that can suggest the real age of the base of the 1st FAU, but the abrupt increase in sediment grain size at the base of the 1st FAU corresponds to the grain size increase that occurred, after the sedimentation of thinner Early Pleistocene sediments, in the Po plain and around the Alps. Studies carried out by Muttoni et al. (2007) demonstrate that at the mouths of the Alpine valleys in Lombardy (east of the study area), a sharp grain size change occurred during the MIS 22 (starting ca. 870 ka BP). Moreover, Valla et al. (2011) have reported that around 800-900 ka BP, in the French Alps, erosion phases stronger than before started. The increased erosion probably produced an increase in the volume and grain size of the debris carried by rivers in the plains surrounding the Alps.

Since the grain size increase is a general trend, and its age agrees with the chronological framework of the 1st FAU, we assume that the sharp transition from fluvial -lacustrine to glaciofluvial sediments occurred during the MIS 22.

The sections reported in Fig. 7 show that the 1st FAU sediments are not in contact with the Lucedio Fault. However, the sediment dip changes nearing the fault. In the northernmost area, the 1st FAU sediments dip towards the south and SE, as one would expect for sediments produced by Alpine rivers flowing from north and NW, while near the Lucedio Fault they dip towards the northern quadrants.

The 2nd FAU lies on an erosion surface carved in the 1st FAU, is mainly of sandy gravel, and reaches a maximum thickness of 20 m. Its top is exposed on the S5 terrace surface ("c" in Fig. 5) and, as discussed above, is covered by the MIS 6 glaciofluvial sediments. The 2nd FAU has been identified also in a borehole made near the Lucedio fault (Fig. 7, section c) embedded between the 1st FAU and the glaciofluvial sediments dated MIS 6. Interbedded within the sandy gravel of the 2nd FAU, a tephra layer has been found. Its age, obtained by the fission track method, is 350±100 ka BP (ENEL, 1984). As sediments are of glaciofluvial origin, the tephra fall could be dated to the MIS 10 (ca. 380-340 ka BP) or MIS 8 (ca. 280-240 ka BP). However, the stratigraphic framework and the tephra age indicate that the tephra was deposited during the MIS 10, because the sediments of the 2nd FAU are older than the deposits forming terrace S4, dated MIS 8. As the 2nd FAU is dated at MIS 10, the erosion surface that separates it from the 1st FAU is probably dated MIS 11.

The 2nd FAU is very near to, or in contact with the Lucedio Fault plane (Fig. 7, section c). However, its deposits are not faulted, but overlie an erosion surface cut into the Early Pleistocene lacustrine and fluvial sediments which are displaced by the fault. This implies that, during the MIS 10, a river was flowing just north of the Lucedio Fault.

The 3rd FAU includes all the sediments younger than that forming the 2nd FAU. The contact between its base and the older sediments occurs through an erosion surface. The older sediments forming this FAU are dated to the MIS 8 in terrace T4.

On the whole, the stratigraphy of the sediment evidenced by boreholes shows that in the area north of the Lucedio Fault sedimentation prevailed, while south of the fault erosion and depositional phases alternated. The difference between the northern and southern areas, however, disappeared during the sedimentation of the 3rd FAU.

4.4. Summary of the stratigraphic data

The combination of morphological and stratigraphic data indicates the chronology of sediments that form the Vercelli Plain.

The deposition of the older glaciofluvial sediments forming the 1st FAU, the highest terraces and part of the S5 surface ("a" and "b" in Fig. 5) started during the MIS 22 (ca. 870 ka BP) and ended before the MIS 11 (ca. 400 ka BP).

The glaciofluvial sediments of the 2nd FAU, forming also part of surface S5 ("c" in Fig. 5), are dated to the MIS 10 (ca. 380-340 ka BP).

The glaciofluvial and lacustrine sediments ("d" in Fig. 5) forming terrace S4 can be dated to the MIS 8 (ca. 280 -240 ka BP).

The glaciofluvial sediments ("e" in Fig. 5) which form a large part of surface S5 and the terraces (5a, 5b, 5c, 5d) in the area near the moraines are dated to the MIS 6 (ca. 180-130 ka BP).

The glaciofluvial sediments ("g" in Fig. 5) that form terrace S7 could have been deposited during the MIS 4 (MIS 4?) but their chronological framework is definitely bracketed between the end of the MIS 6 glacial maximum (ca. 140 ka BP) and the beginning of the MIS 2 (ca. 30 ka BP). Nearly in the same period, or later, in the northern plain, surface L1 was shaped.

The last glaciofluvial sediments ("h" in Fig. 5) forming terrace S8 are dated to the MIS 2 (ca. 30-14 ka BP).

The most recent sediments (i, l, m, in Fig. 5) forming terraces T9, T10, T11 are dated to the late Holocene (ca. 4 ka BP to the present).

5. CONSIDERATIONS ON THE DIFFERENT THICKNESS OF GLACIOFLUVIAL AND FLUVIAL SEDIMENTS

Data show that in the study area the thickness of glaciofluvial and fluvial sediments, dated between 870 ka BP and the Late Holocene, can be quite different (Fig. 8). In general the thickness is greater (reaching 70 m) north of the thrust fronts (which match the Lucedio and Cavourrina faults and the Crescentino and Morano Po flexures). South of the thrust front, the thickness of the glaciofluvial and fluvial sediments is always less than 20 m. In particular, between Fontanetto Po and the area just East of Trino, along the younger terraces, the sediment thickness is less than 10 m.

Because the boreholes are mainly sited in the area between Trino, the RIT and Crescentino, the thickness of the glaciofluvial and fluvial sediments forming the south-eastern Vercelli Plain is less known. In order to evaluate the sediment thickness of the eastern plain, the thickness of the sediments containing phreatic groundwater reported in hydrogeological studies (Regione Piemonte, 2004; Varalda et al., 2006) has been used. In fact, the hydrogeological studies carried out by ENEL (1984) indicated that the base of the phreatic groundwater matches the base of glaciofluvial sediments.

Glaciofluvial sediments are thickest in the plain between Lamporo and the area NW of the RIT (Fig. 5; 7; 8) where the very weathered sediments of the 1st FAU are exposed. In the area NW and W of the RIT, the isopachs are nearly parallel to the Lucedio fault and show that the bottom of the 1st FAU forms two depressions, elongated SW-NE, divided by a small ridge.

The northernmost depression (ND in Fig. 8) lies near the axis of the syncline in the Pliocene and Pleistocene marine sediments, and its depth increases from East to West, in a direction opposite to that of rivers. The southernmost depression (SD in Fig. 8) lies near the Lucedio fault and is closed to the SW. North of Crescentino and near Tricerro, the isopachs are transversal to the Lucedio fault and Crescentino flexure and assume a direction approximately N-S. In the plain east of Trino, the N-S isopachs are nearly parallel to the eastern branch of the Cavourrina fault. Lastly, starting from the area just west of Morano Po and as far as Casale Monferrato, the isopachs have a direction nearly parallel to the flexure and to the northern slope of the Monferrato hills.

Profiles 1 and 2 in Fig. 6 and A-D in Fig. 9 show the variations in the sediment thickness. In the profiles A-D (Fig. 9) it can be observed that there are increases in sediment thickness both west and east of the area between Fontanetto Po and Trino. Moreover, west of Fontanetto Po, the bottom of the sediments dips to the west, therefore in a direction opposite to the slope of the plain. The westward dip of the bottom of the sediments forming the older terraces is greater than that of the most recent deposits. The base of the sediments of terrace S7 lies ca.18 m below that reached east of Fontanetto Po, while the base of the sediments of terrace T8 is 2 m below.

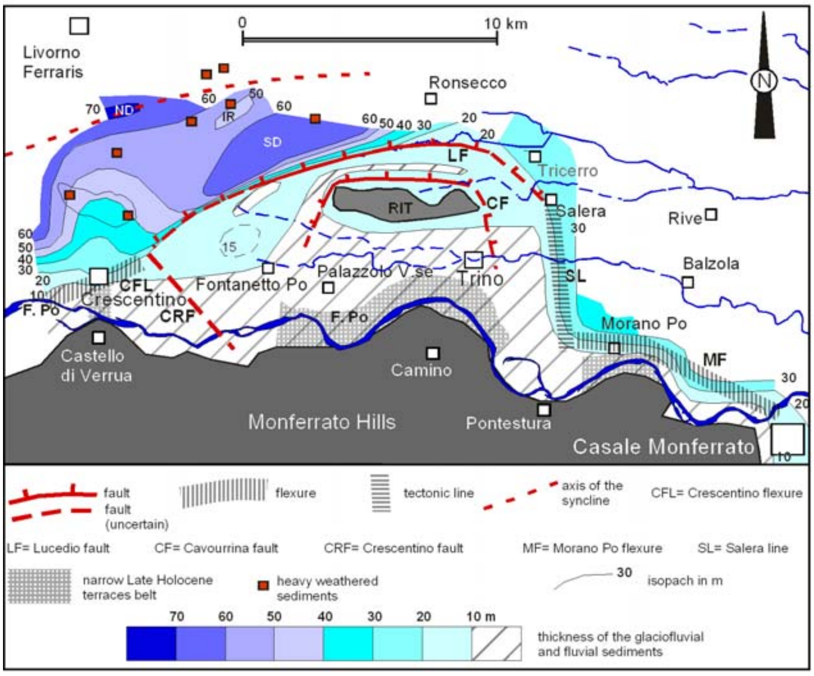

Fig. 8 - Map of the isobaths of the base of glaciofluvial and fluvial sediments and tectonic structures active after the sedimentation of the 1st glaciofluvial and fluvial unit.

6. DATA ON MINOR MORPHOLOGICAL FEATURES

In the Vercelli Plain other secondary morphological evidence can be analyzed and taken into account, such as the change of the inclination of the terrace surfaces. Some observations about small changes of the terrace surfaces, due to the presence of different sedimentary units, were reported above. In this chapter some anomalous slope inclinations observed in the southern part of the plain will be discussed.

Each profile in Fig. 9 refers to the surface of a single terrace. Profile D represents the surface of terrace S7, profile B that of S8, profile A that of terrace T9, and profile C that of surface L2. On the profiles the thickness of the sediments forming the terraces is also reported.

All the profiles show anomalous slope inclination values in the area between Trino and Morano Po (Fig. 9), where the terrace surface inclination is around 0.22 % while in the area west of Trino it is between 0.09 and 0.15% and East of Morano Po between 0.12 and 0.17%.

Between the two gently dipping sectors from the profiles in Fig. 9 a rough value of the scarp height can be estimated. By projecting the slope measured in the western surfaces eastward to the area with the anomaly, it is possible to evaluate the difference between the eastern ground surface and the virtual elevation of the western surface. This method allows us to obtain a height difference of ≥7 m for terrace S7, of ≥6 m for S8, and of about 2 m for terraces T9 and L2.

The anomaly coincides with the area where the glaciofluvial sediments thicken. It is worth noting that all the profiles show that, west of Fontanetto Po, even if the thickness of the sediments increases, no superficial anomaly can be observed.

7. DISCUSSION

Stratigraphy, facies, age and distribution of the sediments and terraces, are useful for the discussion of the geological evolution of the Vercelli Plain starting from the MIS 22. In the first part of the chapter the impact of climatic variations will be considered, and then the stratigraphic data will be analyzed in order to obtain useful information for assessing the tectonic activity that occurred during the last 870 ka.

7.1. The impact of climatic variations and interactions between sedimentation and tectonic activity

The impact of climatic variations on the evolution of the Vercelli Plain is clear, as the plain was formed mainly by glaciofluvial sediments, dated between MIS 10 and MIS 2, connected with the glacial advances and the shaping of the Ivrea moraine amphitheatre. Also the erosion surfaces that separate the different glaciofluvial sedimentary bodies can be ascribed mainly to the glacial retreats that took place during the interglacials. Muttoni et al (2007) suggest that also the grain size increase of the sediments dated to the MIS 22 was caused by climatic variations.

Nevertheless the evolution of the sedimentary units

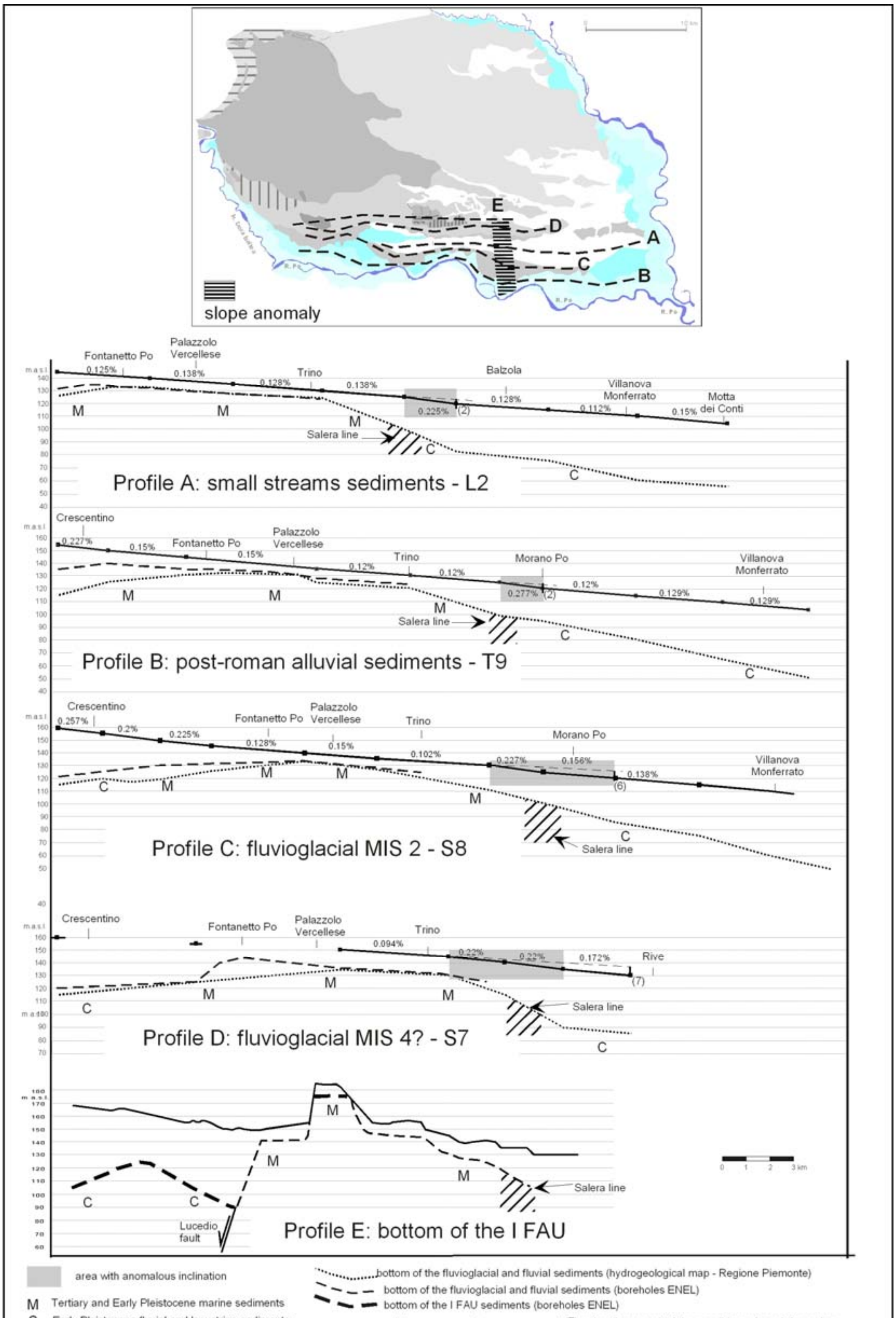

Fig. 9 - Topographic profiles and stratigraphic outlines in the southern Vercelli Plain. The profiles are based on 5 m contour lines reported on the "Carta Tecnica Regionale" at a scale of 1:10.000 in the Piemonte Region. The stratigraphic data are from ENEL (1984) boreholes and from Regione Piemonte (2004).

and the terraces indicates that the climatic impact on the plain evolution changed through time. Starting from MIS 10, the evolution of the plain is characterized by alternating phases of aggradation (glaciofluvial sedimentation) and erosion and pedogenesis. Erosion and pedogenesis occurred between the end of one glacial phase and the start of a new one.

In the Mediterranean area (Macklin et al., 2002), also in places very far from the glaciers, during the interstadials, the rivers incised the sediments deposited during the preceding cold stages, while on the terraces surface the soils evolved.

At least the alternating phases of aggradation and erosion of the late Holocene terraces were climatically driven, as evidence by Giraudi (1998).

Therefore climatic variations have had a great impact on the evolution of the Vercelli Plain during the last 400 ka. For the period preceding the MIS 10, that is during the sedimentation of the 1st FAU, the comparison between the parts of the plain lying north and south of the Lucedio fault shows data that seem to be in contrast with the hypothesis of an evolution driven mainly by the climate.

The plain north of the Lucedio fault is formed by sediments up to 70 m thick, without any evidence of erosion surfaces, heavily weathered at the top, which can be dated between MIS 22 and MIS 12.

South of the fault, the sediments having a similar age have been eroded or form (on the RIT) three different sedimentary bodies whose bases are cut into the tertiary bedrock at progressively lower altitudes. The entire thickness of the three sedimentary units is between 35 and 45 m.

The difference in sedimentation between the areas north and south of the fault cannot be ascribed to climate variations. We believe that, to the north, no strong erosion phases occurred because subsidence prevailed, while to the south the alternating phases of limited aggradation and strong erosion could have been produced by the interaction between tectonic uplift and variations in fluvial regime driven by climate changes.

Michetti et al. (2012), using a boundary R-surface sequence obtained from the stratigraphic data of oil wells, evidenced that the surface (corresponding to the base of the 1st FAU of this paper) is uplifted and eroded at the top of the buried thrust fronts of the Northern Central Po Plain, while in the basins lying outside the fronts, the surface lies below very thick sedimentary cover.

It can be therefore assumed that the shape of the 1st FAU bottom enables the subsidence rate and the distribution and geometry of tectonic structures in the study area to be assessed.

Studies on the Tertiary and Quaternary sediments forming the Monferrato Hills and the basin north of the Monferrato thrust front (Dela Pierre et. al., 1995) show that phases of subsidence of the basin can be correlated to the activity of the thrust front. Mariotti & Doglioni (2000) state that the subsidence in the foredeeps can be ascribed to the load of thrust sheets and/or to the slab pull effect.

The Pleistocene subsidence of the area external to the thrust front is well known and appears to continue the Pliocene trend (SGd'I, 1969a; Bigi et al., 1990; Dela Pierre et al., 1995; Irace et al., 2009).

7.2. Tectonic interpretation based on stratigraphic data

Interpretation of the anomalous changes in thickness and in altitude of the bottom of the 1st FAU and younger sediments makes it possible to evaluate the tectonic activity and the deformation rate of some areas of the plain. Elements showing tectonic deformations connected with the Monferrato thrust front, but also with structures crossing it are discussed below.

7.2.1. Tectonic deformations connected with the Monferrato thrust front

The thickness variations, the dip and the difference in the sedimentation of the 1st FAU are reported in Fig. 7 and Fig. 8. As evidenced above, apart from the RIT, the 1st FAU was not found south of the Lucedio fault and in the Tricerro area, probably because it was eroded. The thrust front, forming the boundary between the two geologically different areas, therefore played a role in the 1st FAU deformation.

The 1st FAU is formed by glaciofluvial sediments of rivers flowing from the northern and western quadrants, but its bottom dips in different directions. In the northernmost area the 1st FAU dips gently to the south. The dip appears regular because the Dora Baltea river flowed from the N and NW. Near the Lucedio fault, the 1st FAU changes its slope, dipping toward the northern quadrants and assuming a direction contrary to the river flow.

The change of the inclination can be ascribed to tectonic deformations. The area where sediments dipping to the north join those dipping to the south (sections of Fig. 7) is the place where sediments reach their maximum thickness.

In the subsidence area W and NW of the RIT, the bottom of 1st FAU sediments (Fig. 8) shows the presence of two depressions (ND and SD) divided by a ridge (IR). The three features have a SW-NE direction nearly parallel to the Lucedio fault.

The northernmost depression (ND) matches the axis of the syncline deforming also the base of the Pliocene marine sediments. The ND bottom cannot have been shaped only by rivers, having also been deformed by tectonic activity. In fact, the base of the 1st FAU dips to the WSW, in a direction opposite to the river flow. For the same reason the bottom of the SD depression, closed to the west, cannot have been shaped only by a river. Above the IR ridge the thickness of the sediments is 15-20 m less that in the ND and SD depressions. The change of the thickness could be explained by a different degree of subsidence.

The deformations of the 1st FAU bottom fit the hypothesis of the presence of tectonic features (probably poorly developed folds) parallel to the Lucedio fault, which can therefore be linked to the activity of the Monferrato thrust front.

The sediments of the 1st FAU do not cross the Lucedio and Cavourrina faults, and so it is difficult to demonstrate when the deformation ended. But if the subsidence implies thrust front activity, then the end of the subsidence could imply the end of the tectonic compression.

The end of the subsidence north of the Lucedio fault is shown by the heavily weathered top of the sediments of the 1st FAU that are exposed near the axis of the subsiding area. In fact, if the subsidence had continued, the (now) weathered sediments would have been completely covered by younger ones. It follows that the subsidence ceased before the end of the 1st FAU sedimentation (MIS 12 - ca. 400 ka BP).

As far as the Cavourrina fault is concerned, it can be observed that its activity ceased before ca. 280-240 ka BP, because undeformed sediments of terrace S4 seal the fault. The fault lies not far from the western, northern and eastern scarps that form the boundaries of the RIT, while no isolated ridge is preserved in the longer and wider area uplifted by the Lucedio fault. The presence of the RIT, and the preservation of the 1st FAU sediments lying on it, suggest that the Cavourrina fault activity produced an uplift stronger than the Lucedio fault. We cannot exclude that the Cavourrina fault activity continued also after 400 ka BP.

Section b in Fig. 7 (in a SSE-NNW direction) shows that the difference in altitude of the base of the 1st FAU sediments between the area where the thickness is greater (ND depression, Fig. 8) and the uplifted area (RIT) can be estimated as about 80-85 m. Nevertheless, the real deformation should be higher because the RIT is situated further down the slope than the ND depression.

Section E of Fig. 9 (direction ca. W-E) shows the base of the 1st FAU sediments between the area west of San Grisante and the RIT. It evidences that the base of the sediment is uplifted near San Grisante, subsiding to the east and uplifting again in the proximity of the Lucedio fault and of the RIT. Fig. 9 shows also that the uplift of the 1st FAU in the San Grisante area occurs in the area north of the fault.

In the area between Morano Po and Casale Monferrato, the thickness of the glaciofluvial and fluvial sediments younger than the MIS 22 increases towards the North and the isopachs lie parallel to the northern Monferrato slope. The trend of the isopachs is similar to that of the area S and SW of Crescentino. The two areas show the same geological characteristics already evidenced by the Tertiary bedrock. The similarity between the two zones enables us to assume that also the isopach trend between Morano Po and Casale Monferrato was produced by a flexure that probably corresponds to a blind thrust front (Fig. 1).

7.2.2. Tectonic deformations crossing the Monferrato thrust front

The trend of the isobaths of the base of the glaciofluvial and fluvial sediments indicates that independent tectonic deformations of the Monferrato front took place.

In fact, in the areas N of Crescentino and NE of Trino, the isopachs assume an approximately N-S direction, crossing the Crescentino flexure and the Lucedio fault (Fig. 6, profiles 1-2; Fig. 8). In particular, in the Crescentino area the isopachs in a N-S direction continue also well to the North of the thrust front. The uplift of the 1st FAU base near San Grisante, pointed out above (section E, Fig. 9), seems to have been produced by N-S deformations.

Both the RIT and terrace S4 north of Crescentino lie in the northern portion of the areas affected by N-S tectonic deformations (Fig. 8). The southern boundaries of the areas affected by these deformations are near Castel Verrua and Camino, where the hills form headlands within the alluvial plain. The N-S deformations, possibly, affect the plain, both north and south of the Lucedio fault and Crescentino flexure, and the Monferrato hills.

On terrace S4, north of Crescentino, the top of the 1st FAU sediments, older than glaciofluvial and palustrine deposits dated to the MIS 8, is strongly pedogenized, but the soil is eroded, at least in part. Probably the deformation phase along the N-S structure started before the sedimentation of the palustrine deposits and favoured the erosion of the soil.

The tectonic deformations along the N-S structure in the Trino area seem to be dated at the MIS 8. In fact, the sedimentary facies of deposits forming the terrace S4 and the thickness of the top sediments (from ca. 4 to 6 m) highlight a decrease of the river energy and the development of a wide marsh on the flood plain, which had never formed before. The changing sedimentary facies fits the hypothesis of difficult drainage to the east, most likely caused by the slow uplift of the area lying NNE or NE of terrace S4.

In the area south of the RIT, the profiles in Fig. 9 show that between Fontanetto Po and the Trino area, to the Monferrato hill slopes, the glaciofluvial and fluvial sediments are less than 10 m thick. West of Fontanetto Po, sediments become thicker and their base dips to the west, that is, against the river flow. Also east of Trino (Fig. 9), in the area where the terrace surfaces show an anomalous slope inclination, sediments become thicker, and the isopachs have a N-S direction.

In the area where the glaciofluvial and fluvial sediments are less than 10-20 m thick the bedrock is of Tertiary marine sediments, while where the thickness exceeds 20 m the bedrock consists of Lower Pleistocene continental sediments. Therefore in the area where the thickness of the glaciofluvial and fluvial sediments is less, the stratigraphic gap is greater due mainly to stronger erosion.

The thickening of sediments to the west and east can be observed also near Tricerro, north of the Lucedio fault (Fig. 6, profiles 1, 2). Whole data suggests that an uplift occurred mainly in the area between Fontanetto Po and Trino, but the area is elongated to the south to the Monferrato Hills, and to the north at least to Tricerro.

A tectonic structure could exist east of the uplifted area. In fact, the increase in the thickness of the glaciofluvial sediments observed east of Trino, evidenced by the isobaths running ca. N-S and crossing the flow direction of rivers, cannot have been produced by rivers alone.

Due to the lack of borehole alignments, it is not clear if the structure is a fault or a flexure. Thus, it can be assumed the presence of an undefined tectonic element (Salera Line), which drove the subsidence of the eastern area. Based of the isopach trend, the Salera Line could also reach Tricerro.

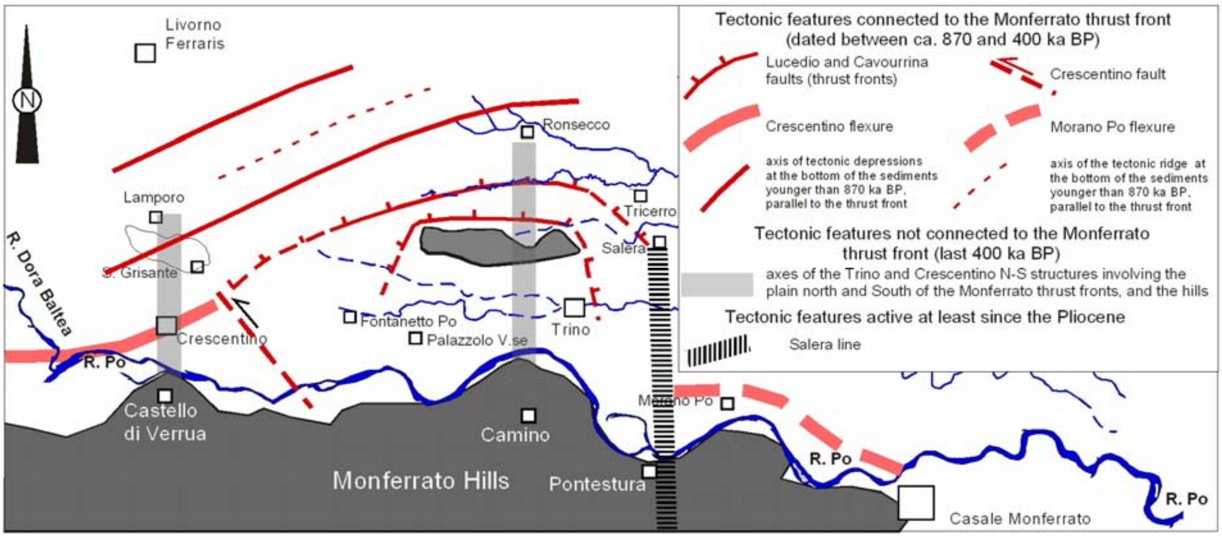

Fig. 10 - Active tectonic structures during the Quaternary.

In the Salera area there is also a flow of saline water from the Tertiary bedrock (ENEL, 1984).

According to the small-scale cartography reported in Irace et al. (2009), along a belt roughly corresponding to the Salera Line, also the isobaths of the Upper Pliocene base are in a direction ca. N-S and the Pliocene sediment thickness increases towards the east.

Faults and lineaments running N-S have been observed in the Monferrato Hills but are not so common. According to Dela Pierre et al. (2003), the paucity of N-S structures in an area having many faults, anticlines and synclines with different directions could be due to their recent (Pliocene and Pleistocene) age.

The Salera Line seems the northern continuation of a strongly asymmetrical syncline lying ca. N-S, at the mouth of the Cerrina Valley in the Po Plain, pointed out by SGd'I (1969a). Furthermore, the hypothetical continuation southwards of the Salera Line would cross the Po river bed in the place where marine sediments are really strongly affected by tectonic deformation.

West of the Salera Line, the Eocene marine sediments, lying below fluvial deposits or outcropping in the Po river bed, reach nearly the latitude of Trino, while east of the Line they reach only the latitude of Morano Po. Therefore, it is possible that the Salera Line produced oblique deformations or that its activity changed in the course of time.

The Lucedio Fault ends eastward against the Salera Line or, alternatively, it could continue in the Salera Line. In other words, this latter might have played a role in the migration northwards of the thrust front and could correspond (although with a different direction) to the transcurrent or transpressive fault hypothesized by Costa (2003) in the same area. Nevertheless, starting at least from MIS 10, the Salera Line seems to have produced only the subsidence of the area East of Trino. To sum up, the Salera Line seems to be a ca. N-S fault with complex kinematics extending from Tricerro to the Cerrina Valley in the Monferrato Hills.

All the structures active during the last 870 ka are reported in Fig. 10.

7.3. Evaluation of the uplift rate

As observed above, the geological and morphological features show strong differences between the uplifted area south of the Lucedio-Cavourrina faults and those north of these. The features are the result of climatic variations and two different phases of tectonic activity.

The maximum difference between the elevation of the 1st FAU base (dated ca. 870 ka BP) in the two areas is more than 80-85 m, and may be estimated as ca. 115 m (Fig. 7, section b); thus, considering that the activity of the Lucedio and Cavourrina faults ended before 400 ka BP, the mean uplift rate of the southern area linked to thrust front activity was 0.25 mm/yr. However in each area differences can be observed in the elevation of the 1st FAU base.

The deformation produced by the N-S uplift area between Trino and Fontanetto Po can be estimated using the variations in the altitude of the base of the glaciofluvial and fluvial sediments where the base dips against upstream, that is, west of Fontanetto Po (Fig. 9). In that area, the base of the sediments dated MIS 4? lies at an elevation 22 m lower than east of Fontanetto Po. As without tectonic deformation the inclination of the base of the sediments should be from west to east, the area between Fontanetto Po and Trino underwent a minimum uplift of 22 m.

Based on the age of the sediments (an interval between 140 and 30 ka BP), the uplift rate is thus >0.16 mm/yr.

If the deformation (more than 12 m) of the base of the sediments dated MIS 2, ca. 30-15 ka old (Fig. 9) is considered, an uplift rate >0.4 mm/yr can be calculated.

The uplift rate of the sediments forming the T9 terrace, dated to the last 2-1.5 ka but whose base could be dated to 3-4 ka, can be obtained using two different data. The first is the value of the dip against upstream, and the second is the anomalous inclination of the terrace surface. The inclination against upstream of the sediment base, west of Fontanetto Po, is about 2 m (Fig. 9) and therefore the uplift rate could be nearly 0.5 mm/yr. The slope inclination anomaly of the S9 surface east of Trino is nearly 2 m in correspondence with the Salera Line, and the uplift rate of the area during the last 2 ka should therefore be nearly 1 mm/yr. However, taking into account the uncertainty of the method, such a rate should be considered the maximum possible.

On the whole, over the last 140 ka, we can estimate an uplift rate between 0.16 and 0.5 mm/yr.

Studies carried out in the Torino Hills (Boano et al., 2004), sited west of the Vercelli Plain, evaluate 1 mm/yr the uplift rate from the Middle Pleistocene to the present.

Within the uplift area between Fontanetto Po and Trino, the southward lowering of the altitude of the base of the glaciofluvial and fluvial sediments and of the terraces surfaces suggests that the northern area is uplifted more than the southern one.

8. TECTONIC INTERPRETATIONS BASED ON MORPHOLOGIC ANOMALIES AND FLUVIAL DIVERSIONS

Studies reported in recent papers on the tectonics of the Po Plain (Burrato et al. 2012a; 2012b; Michetti et al., 2012) have used presumed morphological anomalies and changes in the course of rivers, jointly with structural data, to evaluate the modern tectonic activity and the deformation rates that can be ascribed to structures buried below alluvial sediments.

The epicentral area of the recent seismic crisis (2012) of the Emilian Plain (Galli et al., 2012) matches one of the areas where, according to the previous hypotheses, the drainage was strongly conditioned by the buried tectonic structures. The use of the drainage variations for estimating the activity of structures buried below the alluvial sediments of the Po Plain, therefore, may sometimes give reliable results.

The morphological anomalies, the presence of an anomalous number of terraces and the river diversions, analyzed in this paper, provide the palaeogeographic and chronological data needed in order to complete the observations on the Vercelli Plain presented, in a less detailed way, in some papers (GSQP, 1976; Burrato et al. 2012a; 2012b; Michetti et al., 2012).

The areas with topographic anomalies and complex morphology lie between the axis of the syncline north of the thrust front and the Monferrato Hills (Fig. 2; 4; 5), i.e. the area where the stratigraphic data show the presence of Quaternary tectonic deformations.

The presence of S8 (ca. 30-14 ka BP) south and north of the Po can indicate possible clues of tectonic activity. In fact between Trino and Casale Monferrato, S8 is not far from the Po bed, while NE of Casale Monferrato some small remnants of the S8 terrace are 8-10 km north of the river.

Also south of the Po the S8 terrace remnants suggest an anomaly. The terrace is preserved only west of Verrua Savoia and east of Casale Monferrato. It is likely that the lack of S8 in the intermediate area was caused by the erosion linked to the migration of the river southwards produced by the uplift of the plain north of Crescentino and Trino.

Many river diversions highlighted by the geometry and extent of terraces of the studied area can have been produced by sedimentary and erosion processes, determined by climatic variations or tectonic deformations.

The dominant influence of the climate on the evolution of the plain has already been discussed and shows a strong link between sedimentation and erosion phases, and climatic changes.

Certainly, in the NW Vercelli Plain, the Dora Baltea river bed migrated many times during the last 870 ka, due to the change of shape and extent of the glacial fronts and of the position of the glacial meltwater streams. During the same glacial stages, glacier melting produced more than one meltwater stream fed by different glacial lobes (Carraro et al., 1975; Gianotti et al., 2008).

Some diversions of Alpine rivers and of local streams that took place in the last 2 ka were, or could have been, produced by human impact (rivers Cervo and Sesia in the Vercelli area).

Many river diversions, recognized in the southern Vercelli plain, occurred during the same sedimentation phase and can have been produced by the change in the inclination of the river bed as a consequence of variations in the solid load transport. However most river diversions can have been produced by tectonic deformations along the structures described above.

During the period between MIS 22 and MIS 12 (~0.87-0.4 Ma), the terraces show a diversion to the NW of the river, formed by the Dora Baltea and the DROS, which flowed south of the RIT (Fig. 11a). The diversion is compatible with the uplift of the area south of the Monferrato thrust front, produced by the compression that induced also the deformation of the 1st FAU sediments. In that period, the northern slope of the Monferrato Hills should have been not very far from the RIT.

During the period between MIS 10 and MIS 8 (~0.35 -0.25 Ma), in the Crescentino area, the river changed its course, bypassing San Grisante to the north (Fig. 11b). The age of the diversion is the same as the beginning of the deformation along the N-S structure. During the same period, north of the RIT, a diversion eastwards of the Dora Baltea and the migration of the confluence with the DROS river occurred (Fig. 11b). The diversion may have been produced by the uplift of the area east of Lamporo.

Despite the climatic variations that occurred between MIS 8 and MIS 4?, the river continued to flow north of the Lucedio fault, showing that the climate did not produce river diversions. The river ran parallel to and partly overlapped the thrust front. The position of the river was determined by the presence of the glaciofluvial fan, which limited the lateral variations toward the north, and by the presence of terraces (or of the hillslopes) to the south. The morphological situation was substantially equivalent to that which currently determines the position of the Po from Turin to Crescentino and from Morano to Casale Monferrato.

During MIS 4? (~0.14-0.03 Ma), a series of diversions occurred (Fig. 11c). Initially the river (Baltea +

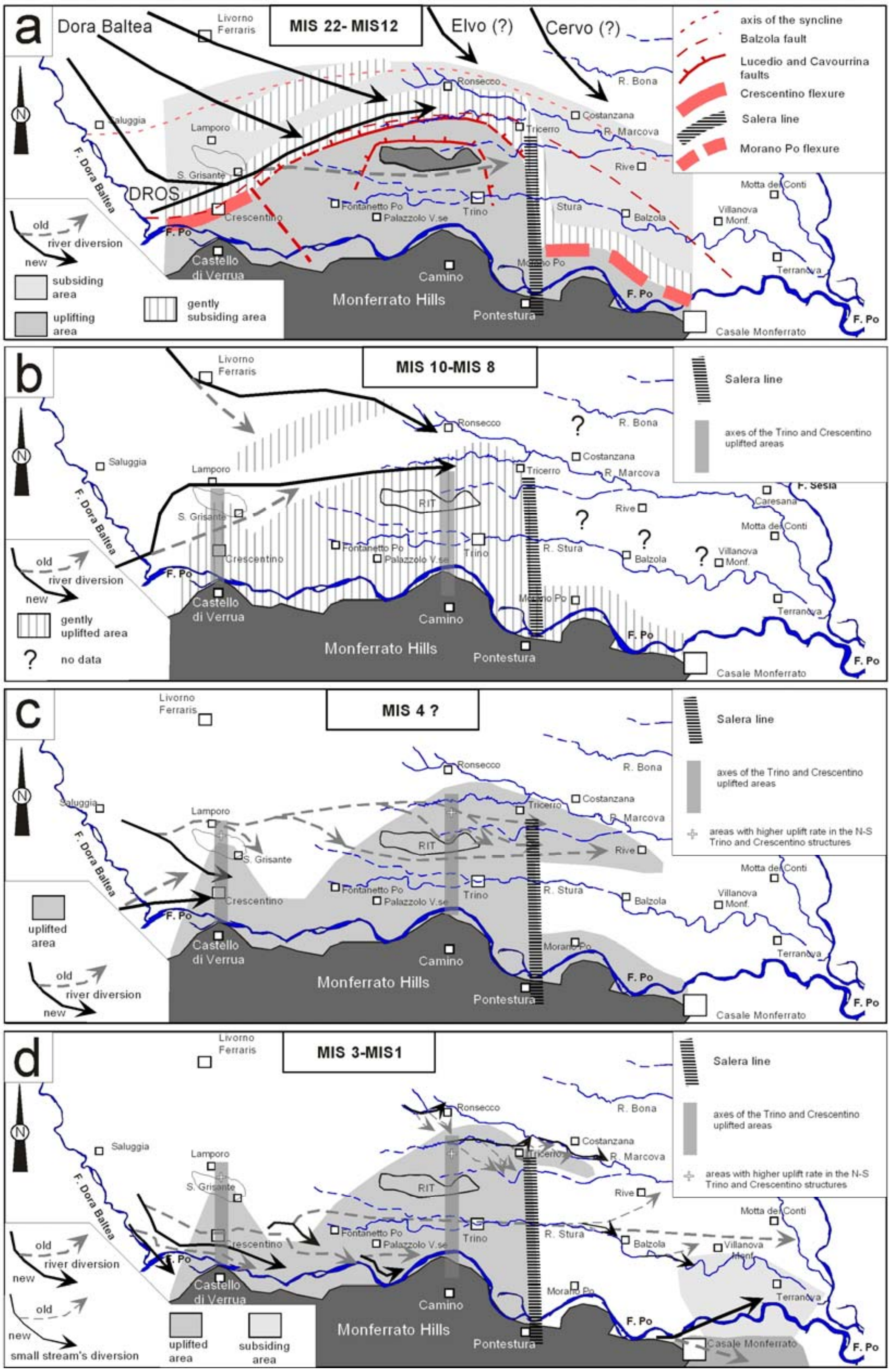

Fig. 11 - Fluvial diversions and tectonic mobility of the Vercelli Plain during the period between MIS 22 and MIS 12 (11A), between MIS 10 and MIS 6 (11B), during MIS 4? (11C), and between MIS 3 and MIS 1 (11D).

DROS), which flowed north of the RIT, migrated slightly towards the south eroding the northern slope of the ridge, then underwent diversions that made it flow to the west and south of the RIT, and later to the west and south of the terrace S4 near San Grisante. These diversions are compatible with the uplift along the N-S structures in the Trino and Crescentino area.

During this period, in the area between Crescentino and Trino an anomalous situation began in the relationship between the position of the river and the Monferrato thrust front, because the river crossed the front and began to flow behind it, showing that the activity of the thrust had ceased.

During the Upper Pleistocene a diversion also occurred of the Po towards the plain north of the Monferrato hills (Carraro, 1976; Carraro et al., 1995), whereas, previously, the river flowed to the south. The Po diversion could have occurred during MIS 4?, just during the phase of continuous diversions of the river in the southern Vercelli Plain.

During MIS 2 - MIS 1 there was a continuous migration to the south of the Po and the Dora Baltea rivers (Fig. 11d) and a significant shift to the west of the confluence of the two rivers.

The river diversions that occurred during the MIS 2 - MIS 1 period are compatible with the uplift of the structures, with a N-S axis, of the Crescentino and Trino areas.

The southward river migration, which produced a strong and continuous erosion of the northern hillslopes, implies that in the Crescentino and Trino uplifted areas, the uplift of the northernmost portion is greater than that of the southern one, confirming the interpretation based on the variations of the altitude of the fluvial sediment base.

The extreme north-south reduction of the Po river drainage basin in the stretch between Crescentino and Trino, described above (Fig. 2), may be a consequence of the southward migration of rivers. In fact, by eroding the hillside they reduced the extent of the basins of the streams that drain into the Po.

In the area between Fontanetto Po and Trino, terraces T9, T10, and T11 form a strip narrower that in the area upstream and downstream. This may be a consequence of the late-Holocene uplift of the area, suggested by the deformation of the fluvial sediment base (Fig. 9).

In the Casale Monferrato area, after the formation of terrace S8, the Po underwent a diversion to the north (Fig. 11d). West of the town, the migration of the river from north to south can be explained by the greater uplift of the northern area of the Trino and Crescentino north-south structures; consequently it can be assumed that the northwards diversion of the river was caused by the uplift of the southern plain and/or by the subsidence of the plain north of the Po.

In the Rive and Balzola area, some small-scale morphological clues on the late Holocene terraces could suggest stream diversions. The natural origin of the diversions is uncertain, but, if the diversions had been natural, it could have been linked to the subsidence of the area NE of Casale Monferrato.

The small valleys (forming surface L2) cutting the river terraces show that during the MIS 2 - MIS 1 period, also the local streams suffered many diversions (Fig. 11d). In particular, in the area between Ronsecco and Tricerro, there are remnants of small valleys abandoned because the streams were diverted N and NE. The diversion may have been produced by the uplift of the northernmost part of the N-S structure.

In the area between Tricerro and Costanzana some morphological features linked to the northward migration of the Marcova stream have been observed. The diversions may have been produced by the uplift of the N-S structure, but also by the uplift of a narrow area (Fig. 11d) parallel to the buried Monferrato front.

9. RELATIONSHIP BETWEEN TECTONIC DEFORMATIONS

In the Arca and Beretta (1985) paper, the uplift rate of some places located near Casale Monferrato and the Vercelli plain is reported. The uplift rate was obtained by means of the comparison between data from geodetic measurements carried out in 1897 and in 1957 AD.

In particular, the points lying along a section from Casale Monferrato to Vercelli, that cross the Monferrato thrust front, indicate an uplift rate of 4.32 mm/yr in the area south of the front, 1.37 mm/yr at Villanova Monferrato, located north of the Po, between the front and the syncline axis, 1.67 mm/yr at Stroppiana and 0.62 mm/yr at Vercelli, located north of the syncline axis. The area south of the Monferrato front has an uplift rate four times greater than in the area north of it.

However, the section does not allow a three-dimensional interpretation and it is therefore not possible to determine whether the deformation occurred exactly on the flexure or on the front, or along other structures transversal to the front.

The morphological features that suggest a possible tectonic activity in the plain near Casale Monferrato are dated to the Upper Pleistocene. No clues showing older deformations have been found. The age of the tectonic activity makes it unlikely that the uplift was produced by the Monferrato thrust front.

10. SUMMARY OF TECTONIC INTERPRETATIONS, CONSIDERATIONS ON SEISMICITY AND ON ASEISMIC MOVEMENTS IN THE AREA

The stratigraphic and morphological characteristics of the Vercelli Plain allow us to assess the evolution during two periods, the first between 870 and ca. 400 ka BP, the second following 400 ka BP.

In the older period, evolution was influenced by the subsidence of the axial area of the syncline lying north of the Lucedio fault and of the Crescentino and Morano Po flexures, while the area between the syncline and the fault and the flexures was less subsiding, and the area south of the fault and the flexures was subject to uplift.

The Lucedio fault was active until ca. 400 ka BP while the Cavourrina fault could have been active also later, but before 280 ka BP. As discussed above, these faults constitute the thrusts front, and therefore we can conclude that the main compression in the Monferrato front ended before or about 400 ka BP.

During the thrust activity the Salera Line and the Crescentino fault acted as transcurrents or transpressives, according to the Costa (2003) hypothesis.

Because the thrust fronts are reported on maps at a scale lower than that of the maps in this paper, and there is uncertainty about their exact position (Fig. 1), it is difficult to establish in detail whether the Crescentino and Morano flexures lie on stretches of the Monferrato thrust front, but it is likely that the flexures are superficial effects produced by the blind thrust.

Interpretation of the data suggests that the fronts corresponding to the Lucedio and Cavourrina faults and to the Crescentino and Morano Po flexures were active at least in the period between 870 and 400 ka BP.

The deformation of the sediments shows that the tectonic compression acted in different ways on the various stretches of the front during the same periods. The dip of the sediments, the degree of deformation and the dip of the fault planes indicate a clear difference between the fronts corresponding to the Lucedio and Cavourrina faults (the farthest from the Monferrato Hills) and the front that produced the flexures.

According to the published papers (SGd'I, 1969a; SGd'I, 1969b; Pieri and Groppi, 1981; Cassano et al., 1985; Dela Pierre et al., 2003a; Festa et al., 2009), excluding the area of the Lucedio and Cavourrina faults, the buried thrust front of the Monferrato-Torino Hills is always near the northern hill slope. Moreover, the Tertiary and Quaternary sediments lying outside the thrust front show an inclination and dip very similar to that observed in the Crescentino flexure.

It can be deduced that, although framed within the activity of the Monferrato thrust front, the Lucedio and Cavourrina faults and the other associated tectonic structures (Crescentino fault and Salera Line) acted in a different way.

The anomalous tectonic evolution of this part of the thrust front started well before the Quaternary. In fact, the Balzola Fault (Fig. 4; 5), which is a lower Pliocene thrust front lying north of the Lucedio fault (SGd'I, 1969a; Pieri & Groppi, 1981; Cassano et al., 1985; Bigi et al., 1990; ENEL, 1994; Dela Pierre et al., 2003a), is nearly parallel to the younger thrust front. Naturally the data on the Quaternary evolution are related to a short period, compared to the length of the tectonic phase, and cannot properly explain the tectonic anomaly of this part of the Monferrato thrust front.

In the last 400 ka, the only possible tectonic motions affect N-S trending structures near Crescentino and Trino. The activity of these structures became stronger during the last 250 ka. According to the deformation of the base of the sediments dated at the last 140 ka, the uplift rate of the north-south Trino structure was between 0.16 and 0.5 mm/yr and the northern part of this structure uplifted more than the southern part.

The N-S deformations were not found in other places north of the Collina di Torino-Monferrato. It follows that these deformations are anomalous, but correspond to an area of anomaly since the lower Pliocene. As the Lucedio and Cavourrina faults are sealed by Middle Pleistocene sediments, the N-S deformations can hardly be connected with the thrust front activity.

Is not clear whether the north-south uplifted areas are included between north-south striking faults or flexures, but east of Trino lies the Salera Line, a very important structure with a long and complex activity. The Line seems to continue from Tricerro to the Cerrina Valley, inside the Monferrato Hills, reaching a length of ca 20 km.

We cannot completely exclude the uplift, since MIS 4?, of a short, narrow area east of Tricerro (Fig. 11c;d) parallel, but external, to the Lower Pliocene Monferrato thrust front.

The seismic catalogues show, for the studied area, the absence of any earthquakes between 1000 to 2006 AD (Rovida et al., 2011; RSNI, 2014a) and the presence, during the period covered by instrumental record, of only four events of ML $\geq$2 (RSNI, 2014 b) located in Trino, Camino and Gabiano, but the depth of the epicentres of two events near Trino has been assessed as more than 50 km below ground level.

Because the area of Casale Monferrato, according to the Arca and Beretta (1985) geodetic data, was affected between 1897 and 1957 AD by vertical movements with values greater than 4 mm / yr $^{-1}$ but there were no earthquakes, one must conclude that the area is characterized by significant aseismic movements. The aseismic uplift rate of Casale Monferrato is at least eight time greater than the maximum rate that can be assumed for the Trino area on the basis of the geological data.

11. CONCLUSIONS

The study of the Quaternary geological evolution of the Vercelli plain shows complex morphological, stratigraphic and tectonic variations. With reference to the main aim of the work, i.e. the validation of the seismotectonic interpretations and the associated seismic hazard implications made by various Authors, we can observe the following.

- Only some stretches of the buried fronts, which were reported or assumed as active by various Authors (Fig. 1), have influenced or induced deformation in the Quaternary sedimentation and drainage. Among the buried thrust-fronts which were active in the past 870 ka (i.e., Lucedio and Cavourrina faults), and the flexures of Crescentino and Morano Po, only the first two produced the most appreciable deformation.
- The Salera Line, along with another possible fault (Crescentino fault), may have played a key role in the northward shift of the thrust fronts of Lucedio and Cavourrina.
- Over the past 400 ka the Monferrato front has no longer been active, or its activity has been very low because the base of the II FAU seals it. Thus, starting at about 400 ka BP, the tectonic evolution of the thrust and of the foredeep basin differed substantially from that of the fronts of the Emilia and Ferrara Folds which are considered active and seismogenic.
- From MIS 10 or before, the Crescentino and Trino structures, transversal to the Monferrato front contributed to the uplift of areas elongated approximately in a NS direction. The uplift affected both the northern

Monferrato slope and areas in the plain south and north of the thrust front. The uplifted Trino area was limited to the east by the Salera Line. There are no indications of structures (faults or flexures) located to the west of the Trino area or forming the limits of the uplifted Crescentino area, although their presence might be assumed.
- It is possible that the aseismic uplift, geodetically recorded in the area of Casale Monferrato, is connected with a structure similar to those of Trino and Crescentino.
- It is possible that there is a structure that produces the uplift of a small area to the east of Tricerro. The structure would lie outside of the oldest Monferrato front, but its trend is sub-parallel to this.
- It is possible that the southern Vercelli Plain was affected by tectonic deformation even in the late Holocene.
- For the period between MIS 4? and the present, the maximum uplift rate of the structure of Trino is evaluated as between 0.16 and 0.5 mm/yr.
- The most important identified structure, that was active also during the Upper Pleistocene and the Holocene, is the Salera Line, an unknown structure with a complex kinematics, which might continue to the south, inside the hills, for a length of about 20 km.

The presence of aseismic uplift (up to 4 mm/yr in the area of Casale Monferrato) in an area with very low seismicity, could justify the deformation of sediments and of terrace surfaces that occurred in the past thousands of years. However, the aseismic uplift, tested only for a few tens of years, may not be significant or may be only one aspect of the long-term tectonic evolution. In addition, as elsewhere in the world, there is the possibility that the area is subject to earthquakes with very long return periods, not reported by the seismic compilations.

The detailed study of the geological evolution of the Vercelli plain has therefore made it possible to characterise the tectonic structures that were active during the Quaternary and to assess the uplift rate.

The recent vertical movements of the surface, however, are not associable with the Monferrato thrust front, as suggested previously, but to north-south unknown structures, transversal to the front, which seem to affect both the plain and the northern portion of the Monferrato hills.

As a concluding remark, it is unlikely that the seismicity of the easternmost Apennine fronts (Emilia and Ferrara Folds) can provide useful information to assess the seismic hazard of the Vercelli Plain and of the Northern Monferrato.

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