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Abstract. The Montello is an elongated hill about 15 km long and 5 km wide located south of the Venetian Alps front and ~ 100 km southwest of Gemona, site of the destructive $M_s \sim 6$, 1976 earthquake sequence. Mio-Pliocene strata in the core of the hill are folded. Seven Quaternary terraces across the western termination of the anticline have also been folded and uplifted. The terraces flank the abandoned Biadene valley, a former course of the Piave river which now flows eastwards along the north side of the hill. Topographic profiles along and transverse to the valley and terraces are used to measure the progressive development of the anticline. Fossil remains and archaeological sites dated with ^{14}C suggest that the Biadene paleovalley was abandoned between 14 and 8 ka (11 ± 3 ka). The successive terraces appear to have been emplaced at the onset of interglacials and interstadials, since about 350 ka. The best fitting terrace ages suggest vertical uplift rates of about 0.5 mm/yr before 172 ka and of about 1 mm/yr after 121 ka. The Montello thus appears to be a growing ramp anticline on top of an active, north dipping thrust that has migrated south of the mountain into the foreland. Modeling the deformation of the terraces as a result of motion on such a thrust ramp requires that it propagated both south and upwards with time but with a constant slip rate (1.8-2 mm/yr). For at least 300 kyr the lateral growth of the anticline kept pushing the course of the Piave river southwestwards, at a rate at first of 10 mm/yr, and then 20 mm/yr. Though the growth rate doubled more than 120 kyr ago, the anticline kept a constant height/length growth ratio ($\simeq 20$) implying self-similar depth/length growth of the thrust underneath. The clustering of historical earthquakes north of Treviso suggests that the thrust responsible for ongoing folding of the Montello slipped seismically three times (778, 1268, 1859 A.D.; intensity $I \geq$ VIII) in the last 2000 years, with events of maximum magnitude close to 6 and with average recurrence time between 500 and 1000 years. NW shortening on NE-SW trending thrusts along the Venetian Alps front is compatible with the direction of convergence between Africa and Europe but does not suffice to absorb this convergence.

1. Introduction

Recent deformation in the Mediterranean has long been interpreted to result from the convergence between Africa and Europe [e.g., *McKenzie*, 1972; *Tapponnier*, 1977; *Anderson and Jackson* 1987; *Westaway et al.* 1990]. In the central Mediterranean (at ≃ 15°E) the rate is ~ 8 mm/yr in a N16°W direction [*De Mets et al.*, 1990] (Figure 1). Instrumental fault plane solutions [*Lyon-Caen*, 1980; *Gasperini et al.*, 1985; *Anderson and Jackson*, 1987; *Nicolas et al.*, 1990; *Fréchet et al.*, 1996; *Frepoli and Amato*, 1997], historical earthquakes (Basel, 1356, I (intensity) ~ X [*Meyer et al.*, 1994], Chamonix, 1905, $I \sim$ IX; Belluno, 1873, $I \sim$ X), and the high topography indicate that such convergence continues in the western Alps and northern Apennines. Nevertheless, while mapping of active thrusts is in progress in other mountain belts [e.g., *Avouac et al.*, 1993; *Tapponnier et al.*, 1990; *Yeats and Lillie*, 1991; *Gaudemer et al.*, 1995; *Meyer et al.*, 1998], few of the faults responsible for the largest historical earthquakes in either range have been identified or characterized. In part, this is because active thrust faulting does not always

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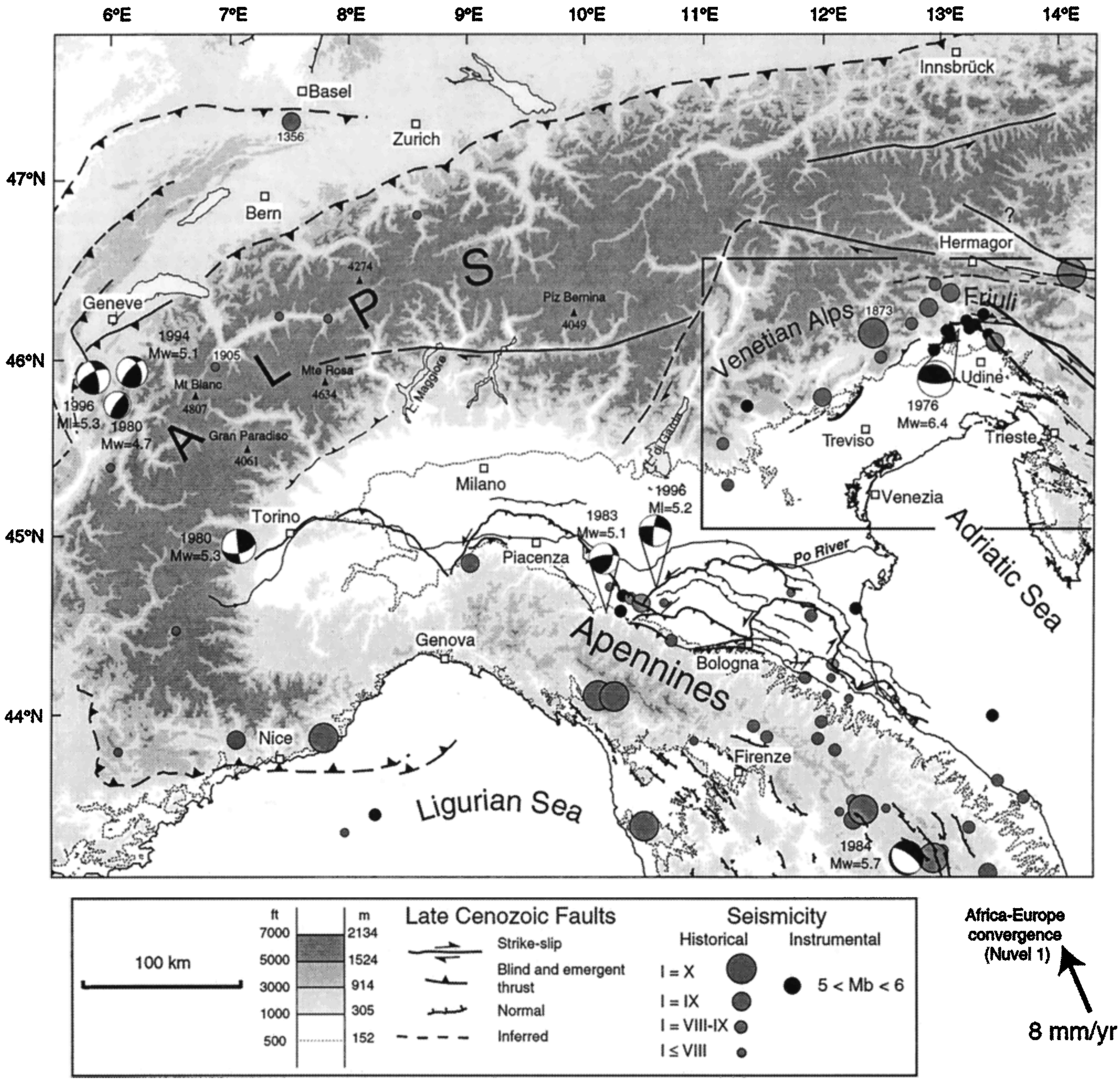


Figure 1. Seismotectonic map of Alps and northern Apennines. Faults are from fieldwork, analysis of Landsat and SPOT images, topographic maps, and, for subsurface, from *Consiglio Nazionale delle Ricerche* [1983]. Historical seismicity is from *Baratta* [1901], *Boschi et al.* [1995] and Monachesi and Stucchi (data, 1997). Instrumental seismicity and earthquake fault-plane solutions are from U.S. Geological Survey, Istituto Nazionale di Geofisica, *Nicolas et al.* [1990], *Fréchet et al.* [1996], and M. Nicolas (personnal communication, 1997) . Topography is from *Defense Mapping Agency* [1988a]. Box indicates location of Figure 2.

induce surface faulting but creates folds, whose ages and growth rates are more difficult to assess [e.g., *King and Vita-Finzi*, 1981; *Philip and Meghraoui*, 1983; *Stein and King*, 1984; *Stein and Yeats*, 1989; *Tapponnier et al.*, 1990; *Avouac et al.*, 1993; *Dolan et al.*, 1995; *Ward and Valensise*, 1996].

In this paper, we examine the relationship between earthquakes, faulting, and folding along the Venetian Alps front (Figure 2), a region frequently shaken by earthquakes. From May to September 1976, for instance, four thrust events with magnitude $\geq$ 5.9 destroyed the northern Friuli villages of Gemona, Osoppo, Madonna, Trasaghis, Magnano, and Ragogna [e.g., *Amato et al.*, 1976; *Bosi et al.*, 1976; *Lyon-Caen*, 1980; *Boschi et al.*, 1995] (Figure 1). Since about 200 A.D., $I \geq$ VIII-isoseismal contours, reconstructed from historical reports of damage [*Baratta*, 1901; *Postpischl*, 1985; *Boschi et al.*, 1995; G. Monachesi and M. Stucchi,

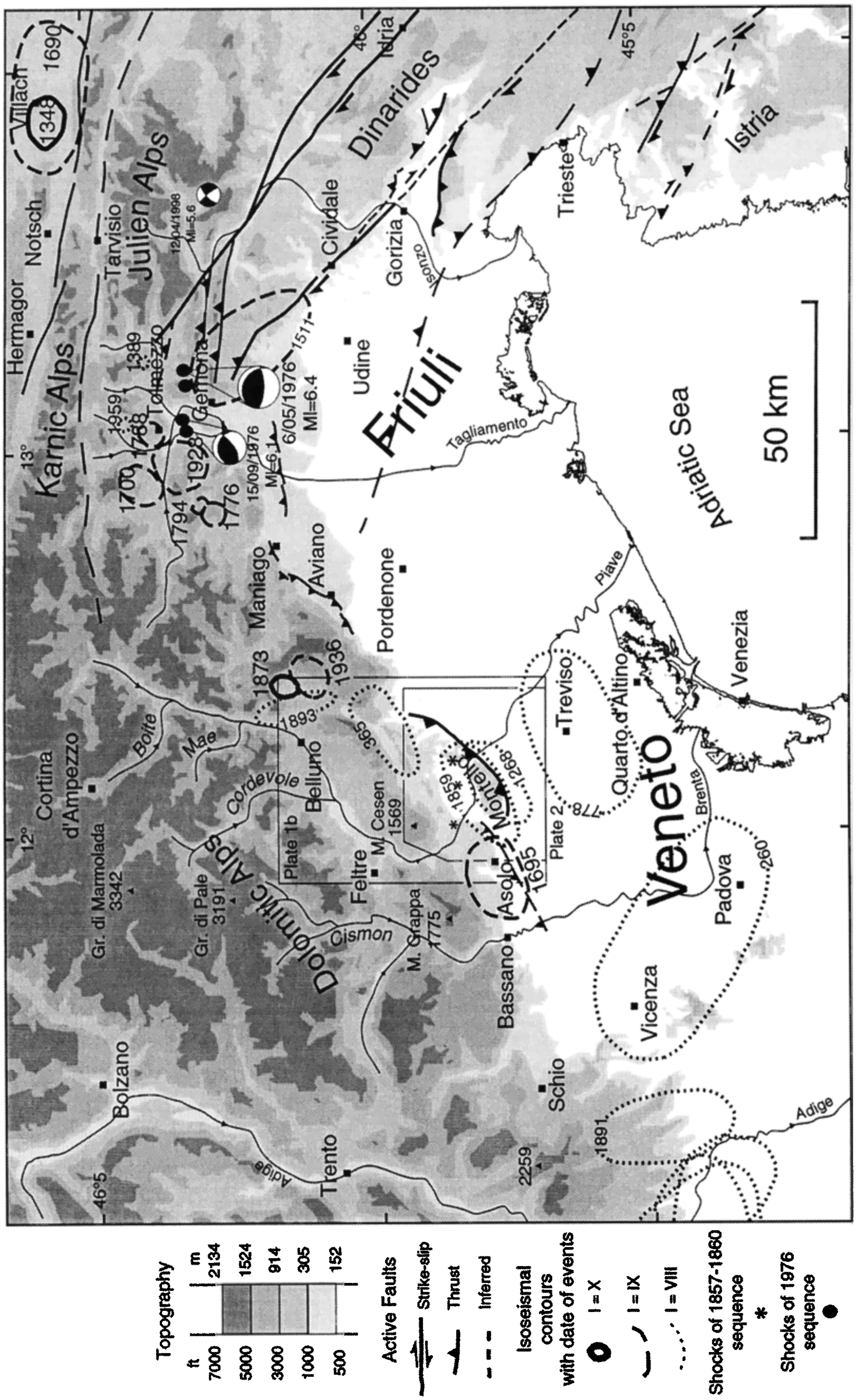


Figure 2. Seismotectonic map of Veneto-Friuli plain and adjacent mountains. Note clustering of historical events north of Udine and Treviso. Topography is from *Defense Mapping Agency* [1988b]. Boxes indicate locations of Plates 1b and 2.

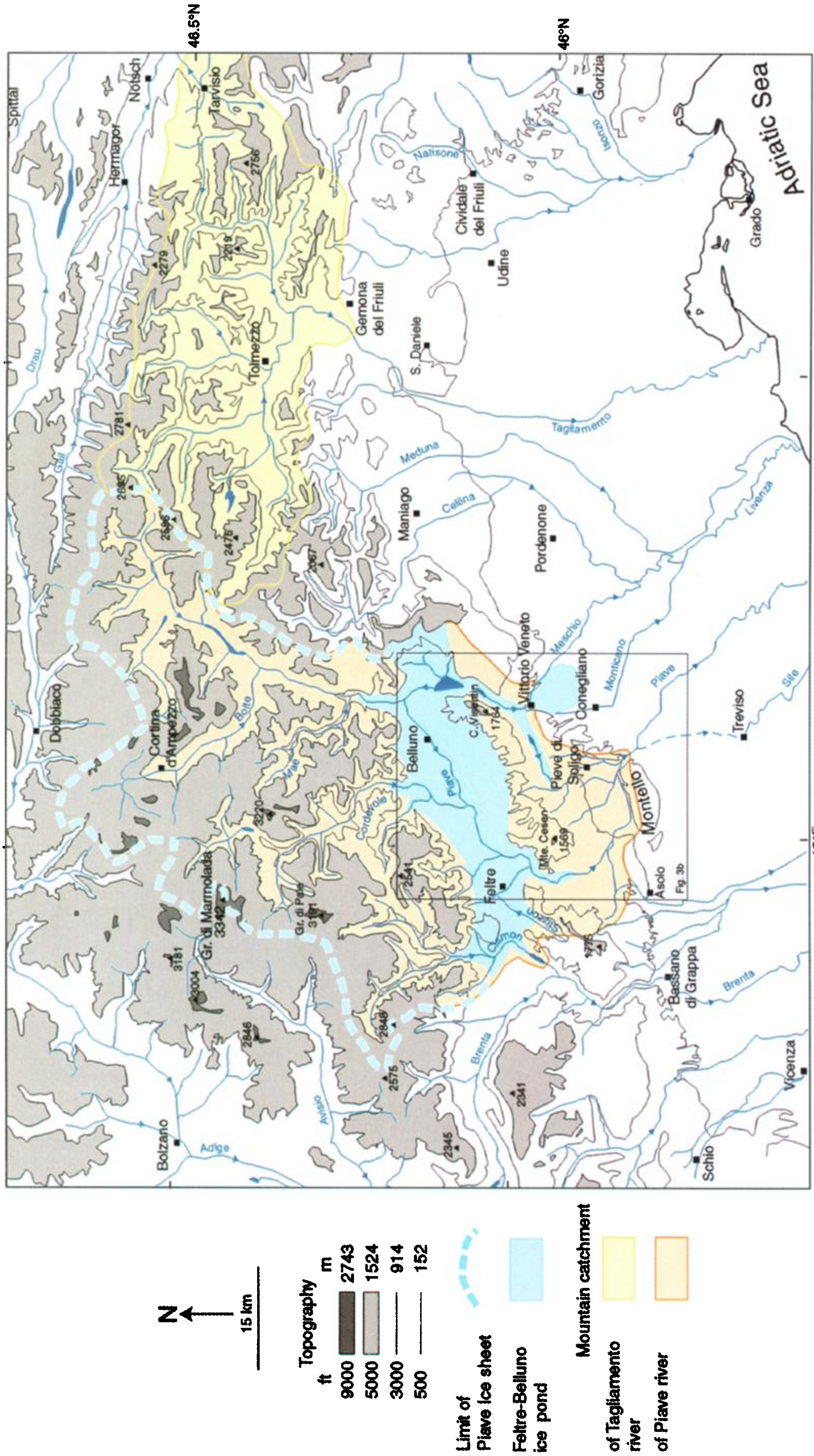


Plate 1a. a) Drainage areas of Piave and Tagliamento. Limit of Piave Würm ice sheet and Feltre-Belluno ice pond are outlined.

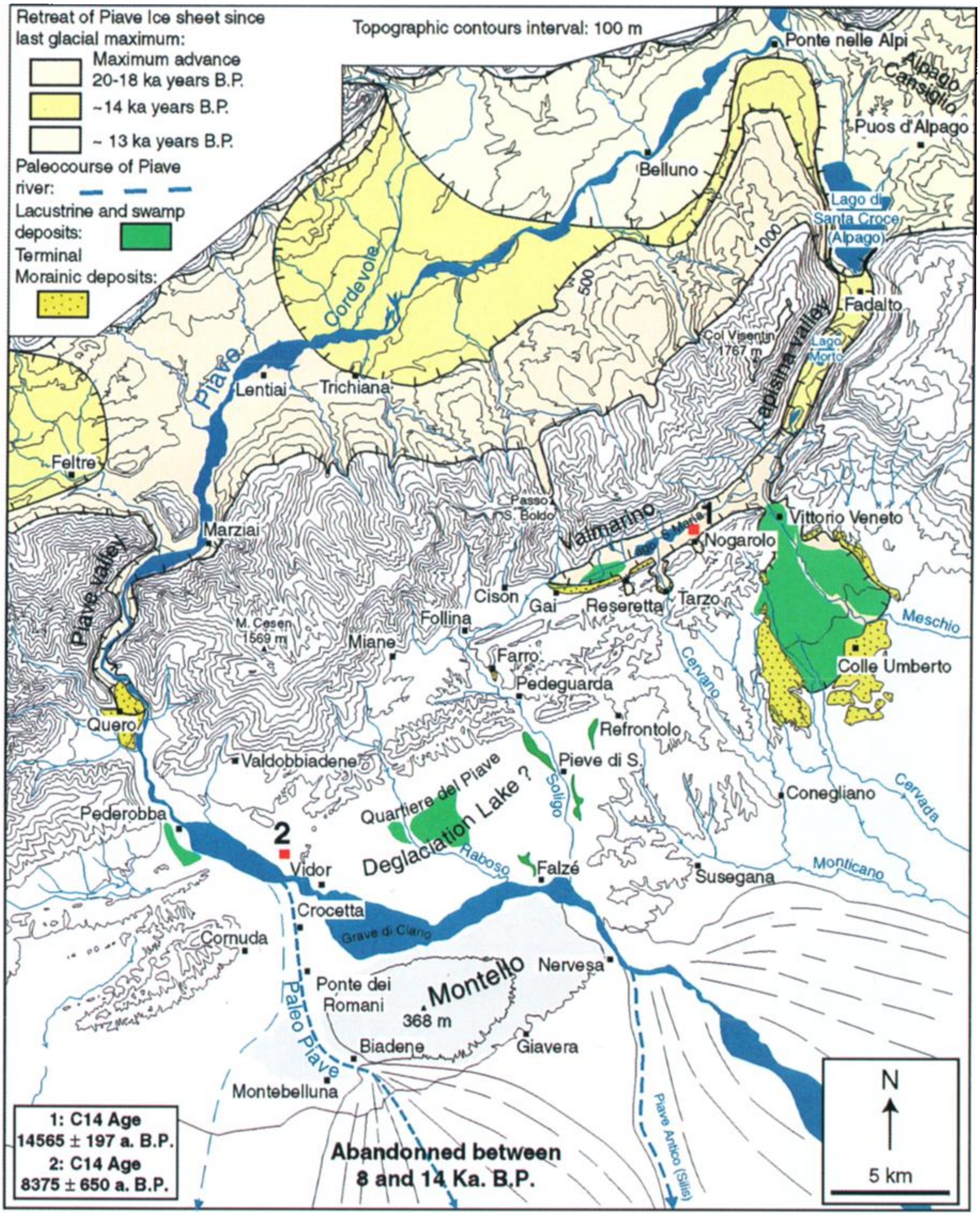


Plate 1b. Würm maximum extent and retreat of Piave ice sheet. Topography is from *Defense Mapping Agency* [1988b]. Terminal moraines and glacial-related deposits are from *Dal Piaz et al.* [1963], *Venzo* [1977] and *Comel* [1955].

DOM4.1, un database di osservazioni macrosismiche di terremoti di area italiana al di sopra della soglia di danno, Grupo Nazionale per la Difesa dei Terremoti, Milan, http://emidius.itim.mi.cnr.it/DOM/home.html, 1997, herein after referred to as Monachesi and Stucchi, data, 1997], reveal two seismicity clusters at the edge of the mountains, one near the outlet of the Tagliamento valley onto the Veneto-Friuli plain and the other ~100 km to southwest, near that of the Piave valley (Figure 2).

In the absence of focal mechanisms, only geodetic or geomorphic evidence can be used to understand the present tectonic regime. Here, we present a study of the region around the Montello, an elongated hill north of Treviso, in which upper Miocene rocks are folded [e.g., *Dal Piaz,* 1942]. We examine the morphology and geology of the Montello fold by combining field observations, satellite image interpretation, and data from topographic and geological maps. The most quantitative evidence comes from a set of uplifted terraces flanking a dry valley near the western end of the fold. Because considerable controversy exists on the timing of deposition and the age of folding of the different conglomerate levels that surround the Montello, a target of detailed study by many geomorphologists and geologists for over 100 years, the first sections of the paper are primarily concerned with establishing the origin and age of these terraces. The ages of the terraces are constrained by combining paleontological and archeological evidence with inferences on climate change without which the fluvial and glacial history of the region cannot be properly understood.

With the geometry and age of the terraces established, it is possible to describe the evolution of the fold for the last 350 kyr and note that the uplift rate apparently changed with time. Two different dislocation models are needed to reproduce the observed terrace forms, which suggests an evolution of the thrust beneath the fold as it developped. Both models, however, require about the same shortening rate. Finally, we discuss the way in which the growth of the Montello anticline is related to the regional tectonics and to historical earthquakes along the foothills of the Venetian Alps.

2. Geology and Geomorphology of the Montello and Adjacent Areas

The Montello is located 10 km south of the Alpine range front north of Treviso and 50 km northwest of the Adriatic sea coast (Figures 1 and 2). The Venetian Alps trend NE-SW and have summits over 3000 m high in the Dolomites. The highest peak, Marmolada (3342 m), is still capped by a glacier (Figures 1 and 2). The southeast border of the mountains rises 1400-2100 m above the Veneto plain (Figure 2 and Figure 3 and Plates 1 and 2) and is bounded by north dipping Tertiary thrusts [*Pieri and Groppi,* 1981; *Doglioni,* 1993]. Alluvial deposition has been rapid in the plain, where rivers flowing from the mountains (Brenta, Piave, Tagliamento, Isonzo) feed large coalescent fans (Figure 2). The surface created by these fans slopes gently (0.1° to 0.2°) from about 130 m above sea level (asl) at the range front down to the Adriatic Sea.

The Montello forms a whaleback-shaped hill, elongated NE-SW, about 15 km long and 5 km wide with a maximum elevation of 368 m, in the upper part of the Veneto plain (Figures 1-3 and Plates 1 and 2). It isolates a small basin (Quartiere del Piave, Figure 3 and Plate 2), now fed by alluvium from the rivers Piave, Soligo, and Raboso, from the rest of the plain. The core of the Montello hill is made of bedded, Mio-Pliocene (chiefly Pontian) conglomerates overlain to the west by Quaternary terrace deposits [e.g., *Dal Piaz,* 1942; *Dal Piaz et al.,* 1963; *Venzo,* 1977]. The Pontian strata dip 10°-30° northwards and southwards on the northern and southern flanks of the hill, respectively, defining a gentle NE-SW trending anticline [e.g., *Dal Piaz,* 1942; *Dal Piaz et al.,* 1963; *Venzo,* 1977], (Figure 3). To the west, north of Montebelluna, the Montello ends in a characteristic periclinal termination (Figure 3 and

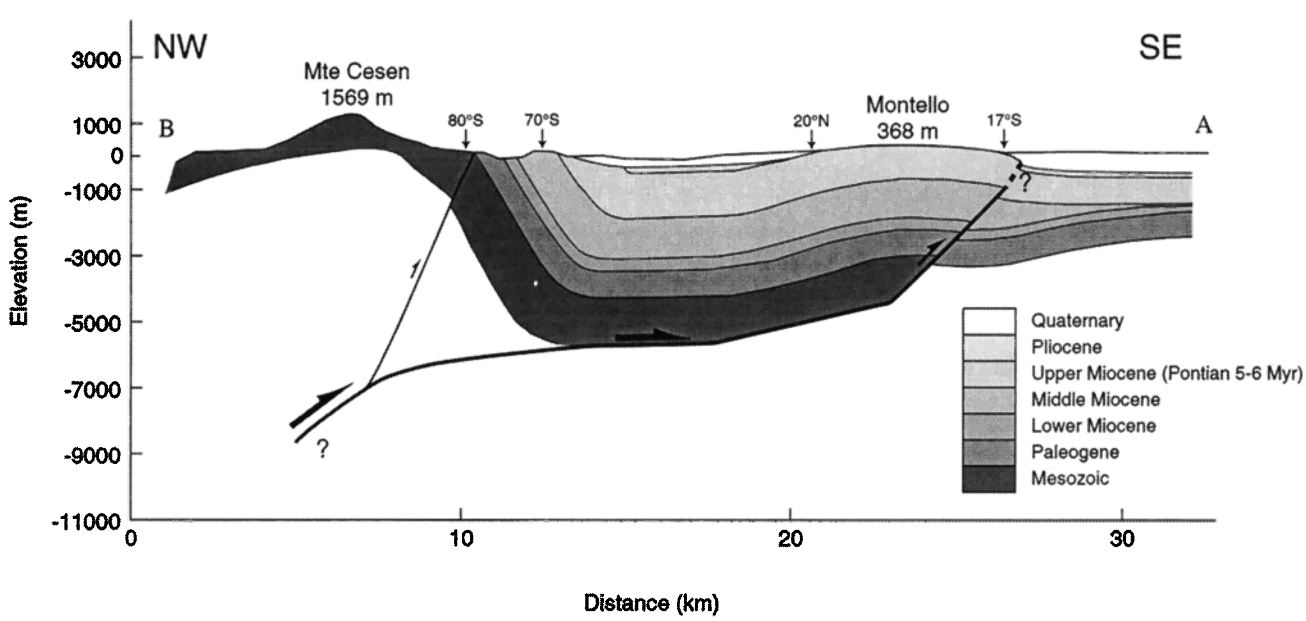


Figure 3. Geological cross section of Montello anticline (see location on Plate 2b) from *Dal Piaz et al.* [1963], *Pieri and Groppi* [1981], *Slejko et al.* [1987] and our observations. Topography is from *IGM* [1950] and is shown with no vertical exaggeration. Bedding dips are from field measurements.

Plate 2). To the northeast, the anticlinal structure continues with a more northerly direction in the hills north of Susegana and Conegliano (Plate 2). The Pontian conglomerate beds of the Montello are exposed again north of the Quartiere del Piave. They lie conformably on a fully developed Miocene sequence on top of reduced Eocene-Oligocene strata. Together these beds make a 50°-70° SE dipping monocline along the base of the mountain front (Plate 2 and Figure 3). The Quartiere del Piave thus forms a "piggy-back," synclinal basin [*Dal Piaz,* 1942] between the Montello and the larger, SE vergent, anticlinal ridge of Mount Cesen to the northwest (Figure 3). Locally, just east of Valdobbiadene, the folded, south dipping Jurassic and Cretaceous limestones of Mount Cesen are thrust over the Oligocene strata. Late Pleistocene deposits in the Quartiere del Piave, which, despite artificial drainage, remains a marshy area, include meters-thick clay horizons, a result of intermittent flooding and swampy to lacustrine deposition.

2.1. Piave Paleovalley

The course of the Piave river forms a remarkable kink along the north side of the Montello (Plates 1 and 2). Flowing southeastwards out of the Venetian Alps near Vidor, it is deviated almost 70° northeastwards along the north flank of the anticline to Falze, over a distance of ~ 10 km. There, the river course veers back towards the southeast, crossing the Montello in a ≃ 500-1000 m wide, NW-SE trending gap or "cluse," the Nervesa strait (Plates 1 and 2). Downstream from Nervesa, the Piave fans out onto the plain and flows southeastwards into the Adriatic sea. In Roman times the Piave course may have followed the western edge of its large fan to Treviso, where it would have merged with the Sile (Silis or Piave Sela) reaching the eastern Laguna di Venezia south of Quarto d'Altino [e.g. *Plinius,* ~60 A.D.; *Comel,* 1955; *Palmieri and Paolillo,* 1993] (Figure 2 and Plates 1 and 2).

No other river crosses the Montello hill, but a dry, concave eastward valley with flattish floor cuts the western termination of the anticline between Ponte dei Romani and Biadene (Figures 4 and 5, Plates 1 and 2 and Plate 3). This dry valley starts south of Crocetta, near where the Piave veers eastwards. It is 700-1200 m wide and roughly at the level of the surrounding alluvial plains, with which it merges north and south of the Montello (Plate 4).

The width and position of the Biadene valley south of the Vidor outlet and the presence, south of Biadene and west of the Giavera, of a large fan that contains granitic and volcanic pebbles derived from the western

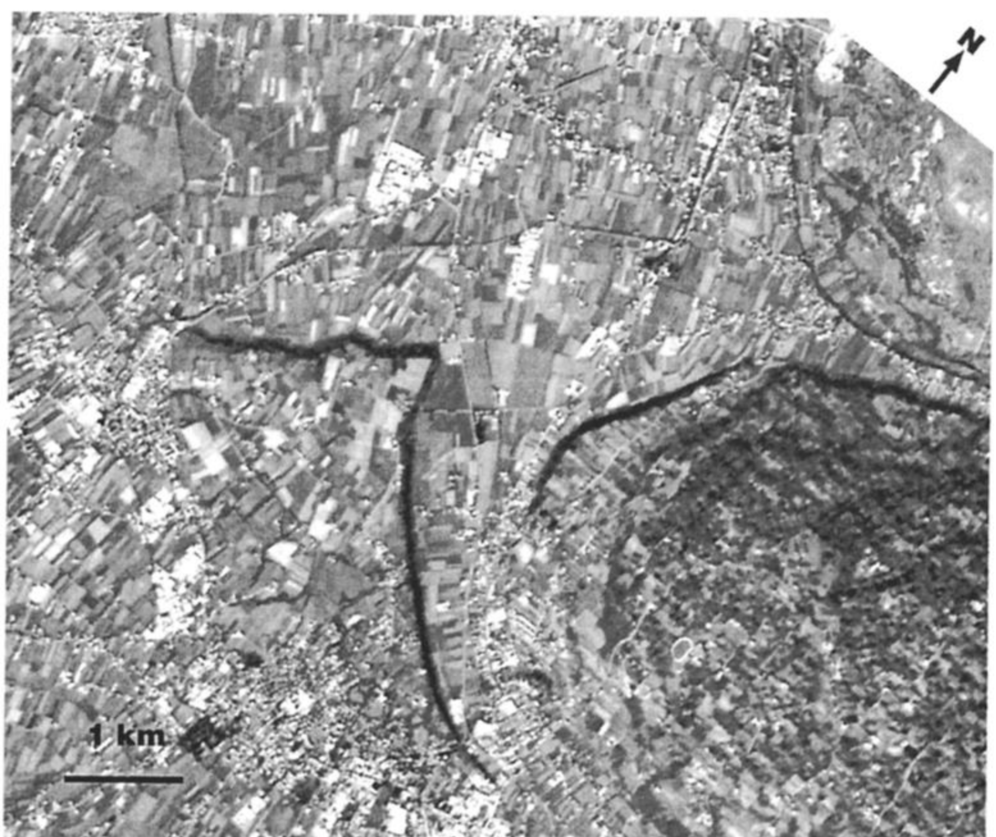


Figure 4. Enlargement of SPOT image of western, periclinal termination of Montello anticline. Note steep risers of Biadene paleovalley and karstic sink holes on high terraces (bottom right).

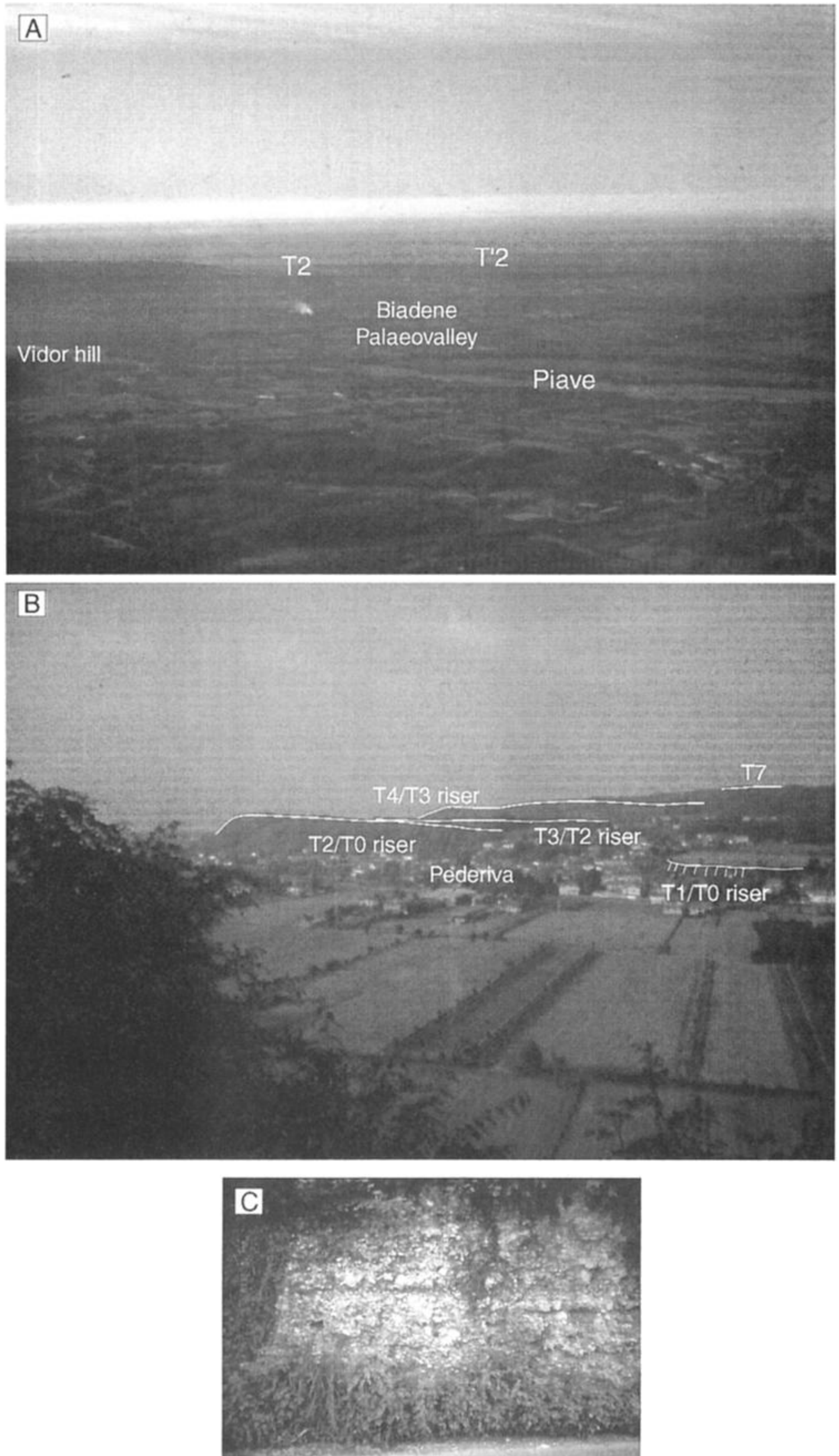


Figure 5. (a) View, toward SW, of Montello's western pericline, and Biadene paleovalley. (b) View, toward NE from Mercato Vecchio, of T1, T2, and T3 risers on east side of abandoned Biadene valley (foreground). (c) Limestone conglomerates of Mercato Vecchio Terrace (T2).

part of the upper catchment of the Piave (Cordevole and Cismon, now a tributary of the Brenta), suggest that this valley originated as a cluse, comparable to that of Nervesa, incised into the western part of the Montello by a former course of the river [*Dal Piaz*, 1942; *Comel*, 1955; *Venzo*, 1977; *Palmieri and Paolillo*, 1993 and references therein] (Plates 1 and 2). Why a change of course of the Piave should have occurred requires explanation in the context of the larger-scale morphological evolution of the region.

2.2. Abandoned, Fluvial Terraces

High terraces flanking the Biadene cluse can be seen on satellite images (Figure 4, Plates 1, 2 and 3) and in the field (Figures 5a-5b), as well as on geological and topographic maps [*Dal Piaz*, 1963]. There are seven inset terraces on the east side of the valley, but only one is unambiguous on the west side (Figure 4 and Plates 2 and 3). The terraces are composed of conglomerates with well rounded pebbles, mostly 5-10 cm in diameter, cemented by a limy matrix. More than 80% of the pebbles are Mesozoic limestones, but there are some volcanic porphyries as well as metamorphic and granitic rocks [*Comel*, 1955; *Venzo*, 1977]. Such pebbles are similar to those found in the paleofan south of Biadene and in the present Piave river [e.g. *Dal Piaz*, 1942; *Comel*, 1955; *Venzo*, 1977]. This suggests that the terraces were emplaced by the Piave river.

The degradation of the terrace surfaces increases with elevation. Here, such degradation is mostly due to the increase in number of karstic sink holes (dollines) due to underground dissolution of the calcareous pebbles and matrix. For the first four terraces east of the abandoned valley (T2 to T5), the sink hole surface density augments regularly (Plate 4b). Though the density of sink holes is greater on T7 than on T6, it is less on either than on T4 and T5. Since these highest terraces are more densely forested, it is possible that vegetation prevents water entering the water table, thus slowing dissolution [*Green and King*, 1996]. It is also possible that T6 and T7 are abrasion surfaces cut in less soluble material or that coalescence of sink holes causes them to be miscounted.

This notwithstanding, the stepwise increase in karstic degradation confirms that the age of the terraces increases with elevation and suggests that they were emplaced over a fairly long time span. *Dal Piaz et al.* [1963] and *Venzo* [1977] have inferred them to be of mid-upper Pleistocene age (between the Riss and the Günz glacials: ~ 120 ka to 400 ka). The increasing age of the terrace surfaces, the provenance of the terrace pebbles, and the generally concave eastwards shape of the terrace risers suggest that such inset terraces on the sides of the dry Biadene meander were abandoned one after the other by the paleo-Piave as it incised the western pericline of the Montello anticline before being deviated eastwards. Since the Piave incises its own fan by at most a few meters south of Nervesa and since high fluvial terraces are found nowhere else downstream from Vidor, where the river enters the most aggradational part of its profile, the preservation of the terraces on either side of the Biadene paleovalley is probably of tectonic origin, as previously suspected [*Comel*, 1955; *Venzo*, 1977].

The existence of more terraces on the east than on the west side of the paleovalley suggests that the former Piave course was being progressively shifted to the west. This makes the present eastward, dogleg deviation of the river along the north flank of the Montello anticline all the more intriguing. One plausible reason for such a change may be found in the glacial history of the region, discussed in section 4.

3. Quantitative Evidence for Quaternary and Ongoing Folding of the Montello

Using one panchromatic SPOT scene (KJ57-259, pixel size 10 m) and topographic maps at scales of 1:25,000 and 1:10,000, with elevation contour intervals of 5 and 1 m, respectively, we mapped the terrace surfaces on either side of the N140-150°E trending Biadene paleovalley. The terraces are labeled T1 to T7, in order of increasing elevation above the paleovalley, which is designated T0 (Figure 4 and Plates 2 and 3). Although, in map view, most of the risers roughly follow the curvature of the valley, three of them, one on the north side of the Montebelluna hill and two on the north side of the Montello, strike at a high angle, roughly parallel to the fold axis and to the present Piave course between Ponte dei Romani and Falze (Figure 4 and Plates 2 and 3). Five of the risers on the north side of the Montello (T2/T0, T3/T2, T4/T3, T5/T4, and T7/T6) have northern segments that trend 30° to 50° more easterly than the paleovalley. This is the form to be expected if the Piave waters had been funnelled into the valley and had been eroding the upstream side of the anticline [e.g. *Meyer et al.*, 1998].

On the map of Plate 3a, it is clear that the risers' bases and tops, particularly where parallel to the paleovalley, intersect the topographic contours, with the higher risers showing greater discordance. This indicates that the terrace surfaces are not horizontal, but are increasingly warped upwards. In order to quantify such warping, noted by *Venzo* [1977] for T2 and T4, we plotted, in the same reference frame, the elevations of points along profiles following the terrace risers' tops and bases and of points spread across the paleovalley floor and on the contiguous alluvial surfaces to the north and south (Plates 4a and 4b). The accuracy of the elevation measurements is that of the 1:10,000 topographic maps.

3.1. Terrace Shapes Projected on NW-SE Sections, Perpendicular to the Anticline Axis

When projected on a plane striking N150°E, approximately parallel to the trend of the paleovalley and perpendicular to the Montello anticline axis (Plates 4a and 4b), the risers' tops and bases profiles have similar shapes. Most of them exhibit 2-3-km-long, rather flat, central segments, some of which slope gently 0.1° to 0.3° towards the northwest (Plate 4a). On the north side of the Montello, such slopes tend to steepen as the risers veer eastwards. The risers are finally truncated by, or merge with, the steep north flank of the anticline, which has been laterally incised by the Piave. The northwards steepening of the slopes tends to be greater for the profiles of the highest terraces. The lengths of the flattish central parts of the profiles also tend to decrease with height. On the south side of the Montello the terrace risers vanish progressively into the smooth south flank of the anticline. On the west side of the paleovalley, the T'2 riser top profile starts sloping gently southeastwards, south of Mercato Vecchio, and steepens markedly just east of Montebelluna (Plate 4a). Such steepening, however, appears to occur in Pontian conglomerates [*Dal Piaz et al.*, 1963], past the southern tip of the T'2 terrace tongue. The southernmost parts of the riser top profiles of T4 and T5 also show clear southeast slopes. The paleovalley floor itself is gently warped above the overall slope (~ 0.4° SE) of the alluvial plains north and south of the Montello. Relative to that slope, taken as a "base level" of deposition along the ancient bed of the Piave, such warping has an amplitude of ~ 11 m and a wavelength about equal to the width of the Montello anticline on section AB (Plates 2 and 4).

Taking the regional, 0.4° SE sloping former depositional base level of the Piave across the Biadene cluse as a reference horizontal axis helps to compare the profile shapes (Plate 4c). The maximum heights of the paleovalley and terraces above this level range from 11 to 200 m (Table 1). Such maxima are reached at the apices of the profiles, where their slopes swing from northwestwards to southeastwards. For the paleovalley and four of the terraces (T2, T'2, T4, T5), the positions of such apices (arrows on Plate 4c) lie within a few hundred meters of each other along the reference axis, closer to the southern than to the northern flank of the Montello anticlinal bulge on section AB. The tops of the two high risers that bound the paleovalley to the east and west have nearly coincident profiles. They thus bound equivalent, probably only slightly diachronous, terrace surfaces (T2, T'2).

3.2. Asymmetry of Terrace Emplacement on NE-SW Sections, Parallel to the Anticline Axis

That the elevations of T2 and T'2 above the paleovalley floor are within a few meters of each other is also clear on profiles transverse to the Biadene paleovalley (Plate 3c). T'2, however, is the only clear terrace level left by the Piave west of that paleovalley. This, the particularly steep slope of the T'2 riser (maximum slope 58°-74°, mean slope 22°-34°), and the contrasting, regular flight of inset terraces on the east side indicate predominant lateral incision to the west and deposition to the east, consistent with unabated westward migration of the Biadene paleoriver meander. The transverse profiles of Plate 3c also imply that entrenchment of the paleo-Piave into the Montello hill probably started shortly after emplacement of terrace T4. Evidence that the Biadene cluse began to form at that time lies in the fact that the riser top of T4 coincides with a marked kink in the average slope of the flank of the Montello east of the cluse. Just below T4, this slope becomes much steeper than that above, which hugs the overall longitudinal profile of the Montello, as extrapolated from the west flank of the Montebelluna hill across the Biadene cluse (Plate 3c). As expected from diffusional degradation of escarpments [e.g. *Hanks et al.*, 1984], the riser slopes tend to be steepest for the lowermost, hence youngest risers (e.g. maximum slope T'2/T0 $\simeq$ 58°-74°), and more gentle for the highest, oldest risers (e.g., maximum slopes T5/T4 $\simeq$ 18°) (Plate 3c). Of the low-level risers, that which bounds the north side of the Montebelluna hill, west of the Biadene cluse, has the gentlest slope (maximum slope 25°-17°), of order of that of the T5/T4 riser, implying that it is much older than the steep Mercato Vecchio riser (Plate 3c). Hence if that riser was cut by the waters of the paleo-Piave, this happened long before the Biadene cluse was abandoned.

Table 1. Degradation, Uplift, and Inferred Ages of Montello's Alluvial Terraces.

Terrace Levels	Sinkhole Density	Uplift, m	Inferred Age, kyr	Slip Rate, mm/yr
T7	~0.098	200	321	1.65
T6	~0.059	161	236	2.12
T5	0.113	146	200	2.09
T4	0.103	130	172	2.2
T3	0.056	104	121	1.78
T2	0.031	70	81	1.77
T1	0	25	29	1.42
Paleovalley T0	0	11	11	1.81

Slip rate on Montello blind thrust is derived from modeling terraces uplifts and shapes.

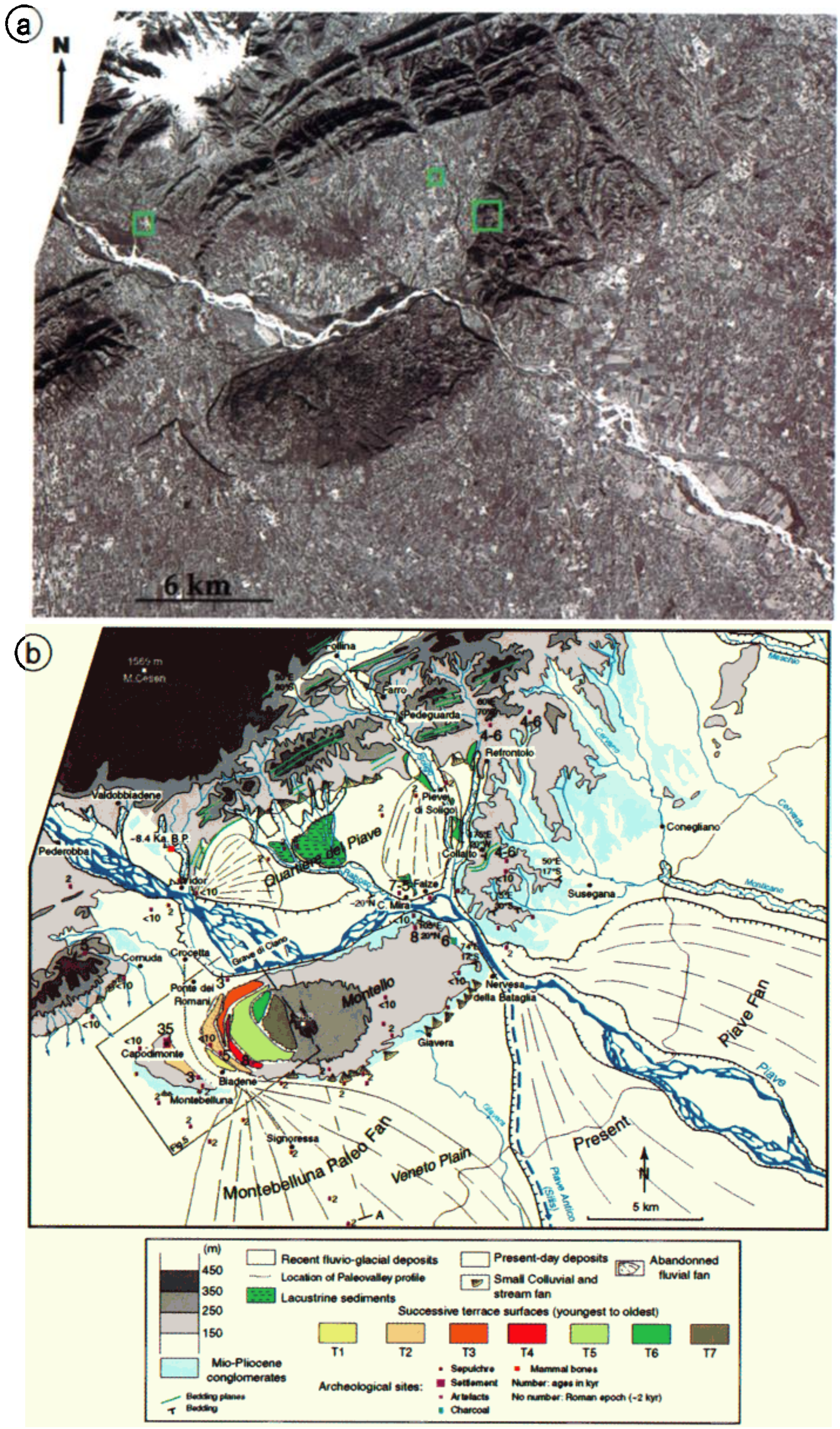


Plate 2. a) Panchromatic SPOT image (KJ 63-257, December 27, 1992) of Montello area (pixel size is 10 m). Note abandoned valley cutting through southwest termination of elongated Montello hill. Green squares are epicenters of shocks during 1857-1860 earthquake sequence; size is proportional to intensity. b) Geomorphic interpretation of Plate 2a. Geology of Quaternary deposits is emphasized. Location of archeological sites near Montello is from *Capuis et al.* [1988]. Numerous Roman epoch sites in Veneto plain south of Montello are not shown on map. Box indicates location of Figure 4 and Plate 3a. Topography is from *Istituto Geografico Militare (IGM)* [1950].

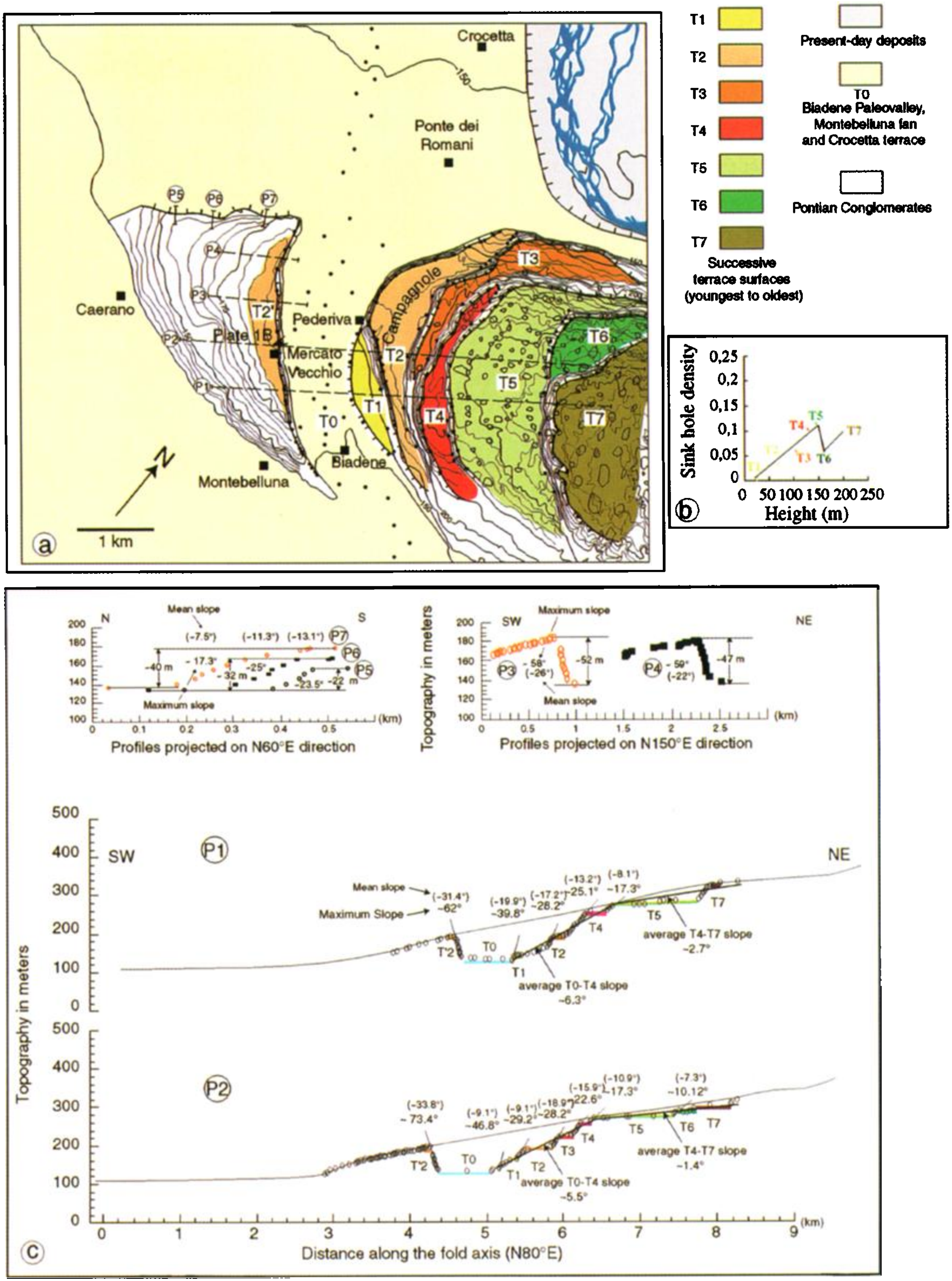


Plate 3. (a) Interpretation of Figure 4, showing uplifted fluvial terraces and recent fluvial deposits. Topography is from *Regione del Veneto* [1992]. Solid circles are topographic trig or selected points and open circles represent sink holes. P1 to P7 are location of profiles transverse to risers. (b) sink hole density versus maximum uplift above T0 for each terrace. Sink hole density is ratio between sink hole surface area (derived from *Regione del Veneto* [1992]) and terrace surface area. (c) Topographic profiles transverse to terrace risers showing shapes of paleovalley incision and mean and maximum slopes of each riser. On P1 and P2, mean riser slope tends to decrease with height. Thin line on P1 and P2 represents overall longitudinal profile of the Montello (extrapolated between T4 and T'2) across the Biadene cluse. Note that terraces above T4 are wider and extend farther towards east.

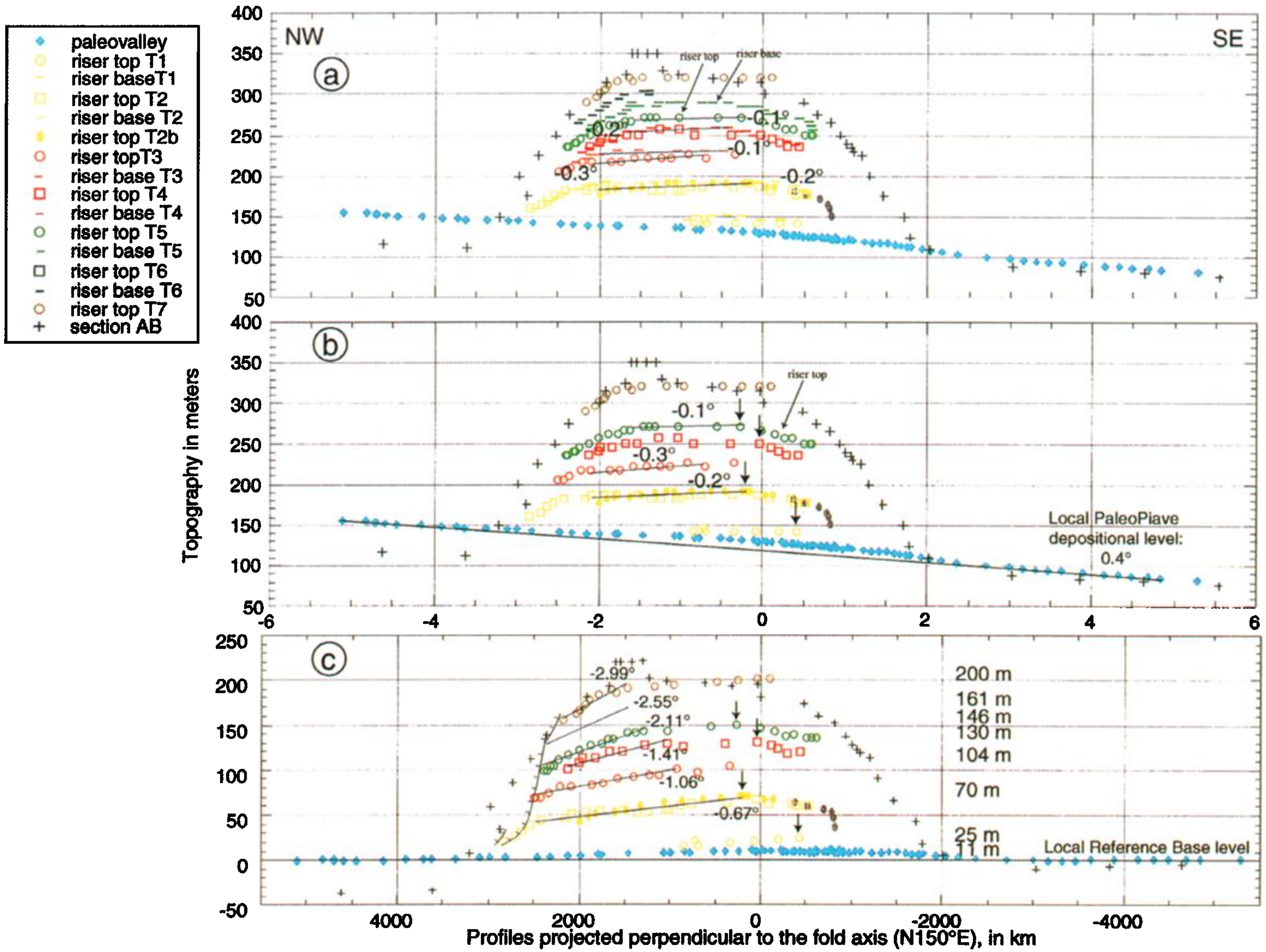


Plate 4. (a) Topographic profiles, perpendicular to Montello anticline axis, along paleovalley floor and bases and tops of terrace risers. Topographic section AB crossing Montello at its summit. Grey symbols are points that do not belong to T'2 surface. (b) Topographic profiles along riser tops and paleovalley floor only. (c) Topographic profiles along riser tops with reference 0.4° slope removed. Vertical exaggeration is 10.

3.3. Progressive Folding and River Entrenchment

Taken together, the different lines of evidence derived from the terrace profiles imply progressive anticlinal folding throughout the time span during which the paleo-Piave flowed around and across the western tip of the Montello, and after the time at which it abandoned the Biadene paleovalley. Assuming that all terraces were initially emplaced with a southeastward slope comparable to the average 0.4°SE slope of the profile between Vidor and Signoressa (Plates 2b and 4b), the present shapes of their surfaces on a section roughly orthogonal to the Montello provide a discrete, quantitative record of such folding at eight different epochs in time.

That three terrace risers' top (T'2, T3, T5) profiles now clearly exhibit segments sloping towards the north ($\simeq$ 0.2°, 0.3°, 0.1°, respectively) opposite to the fluvial deposition slope on the uncorrected profiles (Plate 4b), is unambiguous proof of tectonic warping after deposition. On the section corrected for the regional slope (Plate 4c), the curved shapes of the central 3 km of the profiles provide a measure of folding for each terrace. The more complete profiles of T4 and T5 display more regular folding than the others. The northern segments of six terrace profiles (T2 to T7) have slopes towards the north that steepen with increasing terrace elevation (0.7° to 3°, Plate 4c), hence increasing age, as expected if warped by cumulative tilt increments due to progressive folding. Folding appears to have been asymmetric for all terraces, about an axis shifted towards the southern part of the Montello anticline, as implied by the asymmetric positions of the apices of the profiles. Such asymmetry appears to increase for the younger terraces (T0 to T3), whose profiles tend to have longer, flatter segments uniformly tilted towards the north (Plate 4c). This, and the southward shift of the apices of the profiles (Plate 4c), is consistent with propagation of a thrust ramp towards the surface and the foreland.

Further interpretation of the profiles in terms of folding is limited by the specific circumstances in which the paleo-Piave had to maintain its course past and across the growing anticline. Upstream from the Biadene cluse, as along much of the north side of the Montello, the river waters, either deviated or dammed, then funnelled into the cluse have eroded the north flank of the anticline. North dipping Pontian beds cropping out north of the river bed, near Falze, north of the Nervesa strait, provide clear evidence for such erosion. This has resulted in truncation of the folded terrace profiles by steep risers oriented at a high angle to the paleovalley. Intermediate risers may also exist here, implying that portions of the terraces might have diachronous surfaces. Conversely, as the river fanned out of the cluse past the Biadene narrows, it stopped incising risers into older terraces. Well-defined markers of the deformation thus vanish, as attested by the interruption of most profiles towards the south on Plate 4. This is particularly clear on the east side of the cluse outlet, where loss of incision power was further enhanced by the westwards migration of the river meander. On the other side, by contrast, lateral cutting was maintained by such migration, and we infer the prominent steepening of the top of the Mercato Vecchio riser profile east of Montebelluna to reflect such lateral incision into the south dipping Pontian conglomerates.

In summary, growth of the Montello anticline appears to account for the disposition, shape, and uplift of the terraces along the Biadene paleovalley. Incision of the paleo-Piave into these terraces occurred as a result of folding, while the river maintained, on average, a steady state profile in a part of its course where it would normally be expected to aggrade. Consequently, the uplift of the terraces above this regional profile is a measure of differential tectonic uplift only. Initially, the paleo-Piave probably flowed southeastwards, just west of what is now the summit of the Montello (Plates 2 and 4), in direct continuation with its mountain valley upstream from Vidor. As the anticline grew in both height and length, the river was progressively forced to curve around and past the west tip of the anticline, hugging its periclinal termination. It then became entrenched into this pericline but kept migrating westwards inside the cluse thus formed, due to continued growth of the anticline. The abrupt increase in average slope of the east flank of the Biadene cluse below T4 suggests that the paleoriver meander became entrenched as a rapid increase in uplift rate caused the tip of the fold to propagate faster westwards than further deviation of the paleoriver course could accommodate (Plate 3c).

4. Origin and Age of the Terraces, Relationship With Paleoclimatic Changes

In order to determine the rates of terrace uplift, and the rate of growth in height and length, of the Montello anticline, the terrace ages must be known. Ages reported on the geological maps [Conegliano sheet, scale 1:100,000, *Dal Piaz et al.*, 1963] have generally been based on broad regional correlations of more or less weathered, fluvial, and eolian deposits or soils with the major glacial advances, interglacials and stadials in the Alps [*Venzo*, 1977]. We are not aware of any direct radiometric dating of the terraces or of the Biadene valley floor, but fossil plant debris, mammal bones, and archaeological artefacts have been dated with ^{14}C at several key sites surrounding the Montello. Such ages, and the fluvio-glacial geomorphology, can be used to constrain the regional palaeoclimatic record and concurrent depositional regime of the Piave.

4.1. Maximum advance of the Würm and Riss Glaciers

According to most authors [e.g. *Comel*, 1955; *Venzo*, 1977; *Sacco*, 1899] about 20 to 18 kyr ago at the cli-

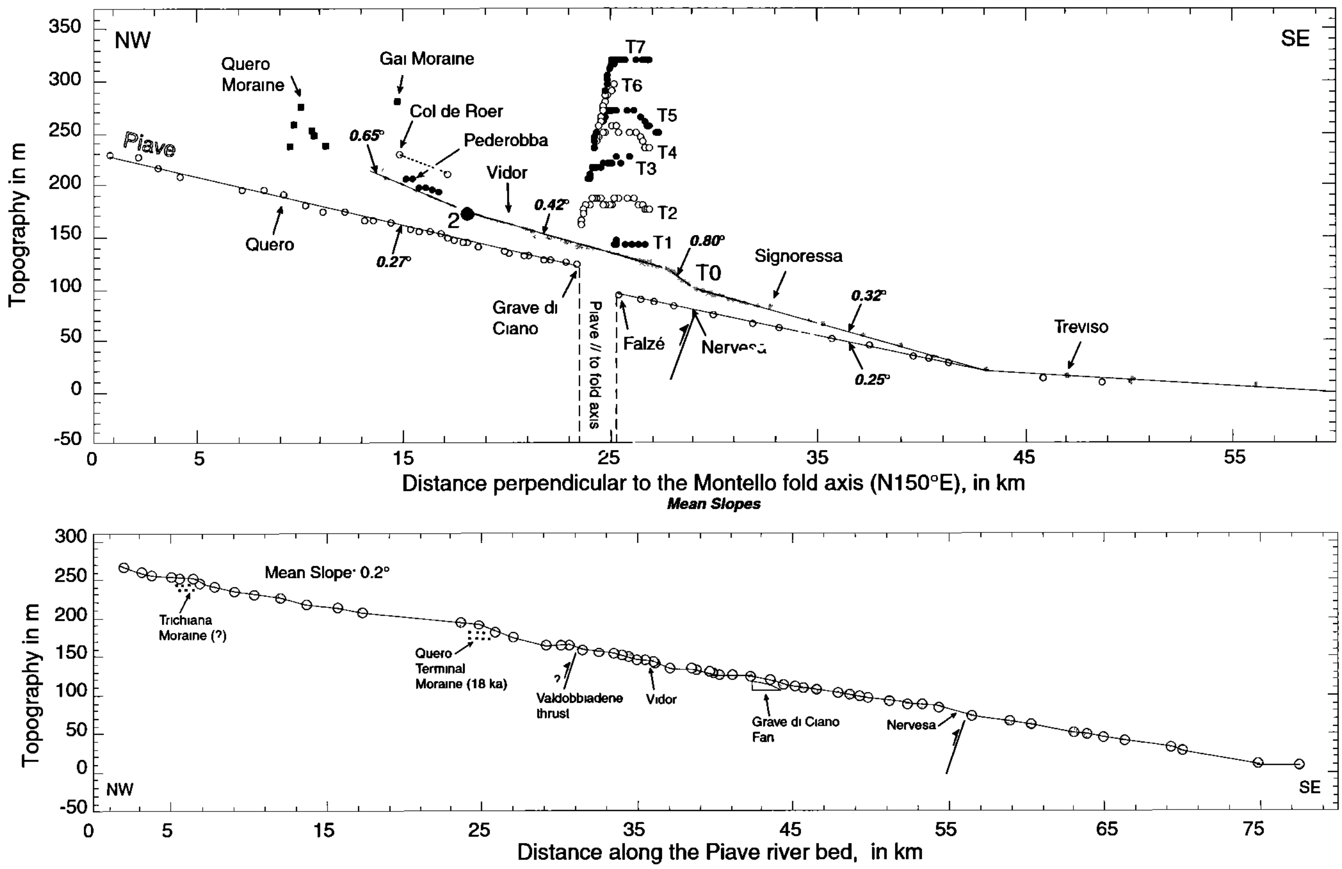


Figure 6. (a) Topographic profiles, projected perpendicular to Montello axis (N150°E), of present-day Piave river, paleovalley, and uplifted terraces (T1 to T7). Note positions of 8.4 ka mammouth bones (2) on paleovalley profile and of LGM Gai and Quero moraines at same altitude. (b) Topographic profile of Piave river bed. Main perturbation along profile is related to Quero moraine.

max of the last glacial age (Last Glacial Maximum (LGM)), the Piave ice sheet, fed by five trunk ice streams spanning the entire catchment of the river in the eastern Dolomites (Cismon, Cordevole, Mae, Boite, Piave) (Plate 1a), filled the Belluno-Feltre synformal trough up to an elevation of ~ 1000 m. This large ice pond, about 15 km wide, 45 km long, and over 600-800 m thick [*Comel,* 1955; *Pellegrini and Zambrano,* 1979] extended southwestwards into the Cismon and Stizzon valleys and spilled over the Marziai and Fadalto narrows, forming two glacier tongues, one along the present Piave valley and the other along the Lapisina valley to the east.

The terminus of the Piave valley glacier was marked by the Quero moraine [*Comel,* 1955; *Venzo,* 1977], whose top now stands at about 280 m asl. This former morainic dam still forms a small perturbation of the otherwise regular, 0.2°-0.3° sloping Piave river profile (Figure 6). The eastern, Lapisino valley glacier, swollen upstream from Fadalto by glaciers from the Alpago-Cansiglio amphitheater (Mounts Cavallo, 2250 m, Messer, 2230 m, and Teverane, 2340 m), forked at the mountain front. Some of the ice spilled southwards onto the plain (~ 100 m asl), forming a pecten pond south of Vittorio Veneto. This pond was rimmed by the large morainic ring of Colle Umberto (top ~ 140 m asl). The main, curved branch of the glacier, structurally guided by the steep Oligocene beds north of Tarzo and Nogarolo, continued straight westwards into the Valmarino. The terminus of that branch, which was perched at an elevation of 225 m asl, ~ 75 m above the Vittorio Veneto branch (Plate 1 and Figure 6), was marked by the moraine ridge of Gai (top ~ 280 m asl). Other small spills breached the Oligocene ridge at the Nogarolo, Tarzo and Reseretta narrows, which now hang ~ 100 m above the valley floor of the Valmarino trunk.

The Riss glaciers probably extended somewhat farther than the morphologically fresher Würm terminal moraines. The Lapisino glacier, for instance, swollen by an ice fall spilling over the San Boldo pass (~ 720 m asl) into the Valmarino, may have reached Pedeguarda (~ 150 m asl), in the Soligo valley. Old, boulder-rich, moraine deposits crop out on the west flank of the Farro hill (~ 250 m asl), ~ 2 km upstream from Pedeguarda [e.g. *Comel,* 1955]. Similarly, about 150 ka ago, the

Piave glacier may have reached Vidor (150-160 m asl) beyond the lateral moraines south of Valdobbiadene and the ~ 220 m asl high terraces of Col de Roer, Riva Milano and San Giacomo [*Dal Piaz,* 1942], on the east side of the Piave river. The pecten pond south of Vittorio Veneto associated with the Riss maximum was probably larger than that of the Würm maximum, as suggested by the existence of small, more degraded, morainic hills south of Colle Umberto.

4.2. Würm Glacial Retreat

The Würm glacial retreat probably started between 15 and 16 ka, with the first pulse of global warming accompanied by a fast rise in sea level [e.g. *Fairbanks,* 1989]. Fossil pine stumps in colluvial sands overlying varved clays near Fornaci di Revine in Valmarino (1, Plate 1b) have yielded ^{14}C ages of 14,765 and 14,370 yr B.P. [*Venzo,* 1977]. This suggests that the Lapisino glacier had already abandoned most of the hanging Valmarino valley by 14.5 ka. The varved clays were probably deposited in a lake fed by glacial meltwater in front of the retreating ice tongue. Similar swamps and lakes related to deglaciation, of which the Santa Maria lakes are the clearest present-day remnants, must have formed in the Valmarino valley and Vittorio Veneto amphitheater upstream from the Gai and Colle Umberto terminal moraines (Plate 1). By 13 ka the ice front in the Lapisina valley had probably withdrawn to the Fadalto moraine [e.g. *Venzo,* 1977] whose elevation ranges between 480 and 580 m asl, 300 m above the Lago Morto downstream. This particularly high and fresh moraine, with big boulders, which dams the large Santa Croce lake (~ 380 m asl) may have marked a point at which the Lapisino glacier stagnated [e.g. *Comel,* 1955] during the Younger Dryas cold spell, until about 11 ka.

The Piave valley glacier probably withdrew in a comparable manner, abandoning the Quero terminal moraine, whose elevation is comparable to that of the Gai moraine, sometime after 16 ka. By 14 ka the Feltre-Belluno ice pond had probably shrunk by half and perhaps divided into two lobes. Possibly, the principal, eastern lobe then extended to the morainic hills of Trichiana (Plate 1). Around 13 ka, the ice front in the Feltre synclinal trough still probably reached south of Belluno and of the Cordevole valley outlet, where young, fresh morainic rings (Marocche), analogous to that at Fadalto and at similar elevation, are particularly prominent [*Dal Piaz,* 1942; *Venzo,* 1977]. At the end of the Younger Dryas (~ 11 ka) the ice sheet inside the Belluno trough vanished completely, and the trunk glaciers in the distinct valleys of the Piave catchment withdrew rapidly to higher elevations [*Venzo,* 1977].

The scenario summarized here is consistent with the overall distribution of high-level moraines observed in the region. In particular, Würm moraines on the north flank of the limestone ridge that bounds the Feltre trough to the south reach elevations of ≻ 1000 m southwest of Belluno and of ~ 700 m north of the San Boldo pass [*Comel,* 1955; *Dall'Arche et al.,* 1979]. It is also compatible with the position of geomorphic slope breaks due to lateral glacial abrasion that are clearly visible on topographic maps and satellite images.

4.3. Post Glacial Aggradation and Holocene Incision, Demise of the Biadene Valley

Postglacial warming, causing rapid melting of the ice in the Lapisina and Feltre valleys and the advent of a more humid climate, resulted in intense flushing of water and debris, particularly by the Piave, the largest permanent river to drain the Belluno trough, and by the Soligo and Meschio (Plates 1 and 2). Such flushing, in turn, triggered widespread deposition downstream from the mountain front, past the river valley outlets onto the piedmont. The formation of a similar late Pleistocene-Holocene bajada due to the coalescence of large fans fed by the melting of south Alpine glaciers is well documented on the north side of the Po plain between Lago Maggiore and Lago di Garda [*Guzzetti et al.,* 1996].

The topographic and geomorphic evidence summarized in Plates 1 and 2 and in Figure 6 suggests that this powerful aggradation episode was responsible for the emplacement, by the Piave and Soligo, of most of the smooth, gently sloping terraces that stand at elevations below ~ 200 m downstream from Pederobba and Cison di Valmarino, between 15 and 10 ka. These terraces, which are now incised by up to 35 m by the two rivers, form a broad, uniform surface, with interior risers only a few meters high, that includes the Biadene paleovalley (yellow T0 surface, west of the Refrontolo-Nervesa line on Plate 2). That the emplacement of these terraces postdates the Last Glacial Maximum (LGM) is consistent with the view of *Comel* [1955], but not with that of other authors such as *Venzo* [1977], who attributed ages older than 50 ka to most of them. Mammal bones (mammouth and megaceros), however, dated at 8375 ± 650 yr B.P. by ^{14}C, have been found in lacustrine clays just north of Fornace di San Giovanni di Valdobbiadene (2, Plate 1b and Plate 2) at a depth of ~ 3-4 m and at an elevation of ~ 165 m. This age was dismissed by *Venzo* [1977] because it contradicted ideas in vogue at that time, prior to the advent of Global Change, but taken at face value, without prejudice, it simply corroborates the young age of the terraces. The bones, whose age slightly postdates the end of deglaciation, are only a few meters below the T0 aggradation surface (see profiles of Figure 6). Also, when extrapolated upstream, the regular profile of T0 merges smoothly with the base of the 18 ka Quero moraine (Figure 6). Such observations concur to establish the postglacial age of T0.

During much of the aggradation episode related to postglacial warming, the Piave thus probably still crossed the Montello through the Biadene cluse. Only after the end of that episode, around the beginning of the Holocene, did it abandon this passage and start incising

T0 to flow eastwards across the Nervesa Strait. The ^{14}C ages of small terrace surfaces in the Upper Piave valley also suggest that the late Pleistocene (15-10 ka) was a period of predominant aggradation while the Holocene was characterized by incision and degradation [*Surian,* 1998]. How the switch from one cluse to the other took place, leading to the east directed dogleg deviation of the river, is not fully understood. However, one clue, hinted at by *Comel* [1955] may lie in the hanging position of the Valmarino glacial tongue.

At the peak of the glacial age the valleys of the Piave and Soligo, coming from the two main valley glaciers north of the Montello, were probably distinct. The Soligo was then the only river to cross the Nervesa Strait. The first pulse of deglaciation, around 15 ka [*Fairbanks,* 1989], probably caused much of the Quartiere del Piave basin to be filled with large amounts of sediment and flooded by water (i.e., by a lake) as suggested by the presence of thick lacustrine clay horizons in its middle part (Plate 1b and 2). Much of the water and sediment discharge probably came from the Piave, given the contribution of meltwater coming from more than 200 km^3 of ice stored in the Feltre-Belluno pond (Plate 1). Meltwaters flowing down the Soligo valley from the retreating Lapisino glacier also contributed at first. But such a flow was suddenly cut off as this glacier withdrew past the entrance of the hanging Valmarino valley, north of Vittorio Veneto (Plate 1b). At this moment the Lapisino glacier meltwaters were diverted into the Meschio, and the Soligo discharge abruptly decreased [*Comel,* 1955]. Later, as deglaciation-related aggradation ceased, around 10 ka, the flat area inundated predominantly by the Piave north of the Montello probably started to drain preferentially through the Nervesa Strait, and the river abandoned the Biadene passage.

This scenario, which may be described as a capture of the upper course of the Piave by the lower course of the Soligo, requires that the two rivers merged through flooding but may have been facilitated by the somewhat deeper valley of the Soligo, downstream from Falze, prior to 15 ka. As befit streams closer to their headwaters, the initial profile of the Soligo might have been steeper. It could therefore have incised somewhat deeper into the regional landscape, despite its small size [e.g. *Bull,* 1991; *Van Der Woerd,* 1998]. In any event, the absence of large and high inset fluvial terraces on either side of the Nervesa Strait is consistent with the inference that it was initially cut by that smaller river, not by the Piave.

To conclude, abandonment of the Biadene paleovalley probably occurred after 14 ka and prior to 8 ka. The absence of palaeolithic artefacts on the surface of T0 [*Capuis et al.,* 1988] supports the idea that it did not predate the beginning of the Holocene by much. We thus take the most likely age of abandonment of the Biadene cluse to be 11 ± 3 ka. Later, the Piave course essentially kept the geometry that it acquired at that time.

4.4. Emplacement and Age of the West Montello Terraces

By contrast, with the recent deviation of the Piave through the Nervesa Strait the seven terraces inset into the west end of the Montello show that the Biadene valley was a long lived passageway of the river across the anticline.

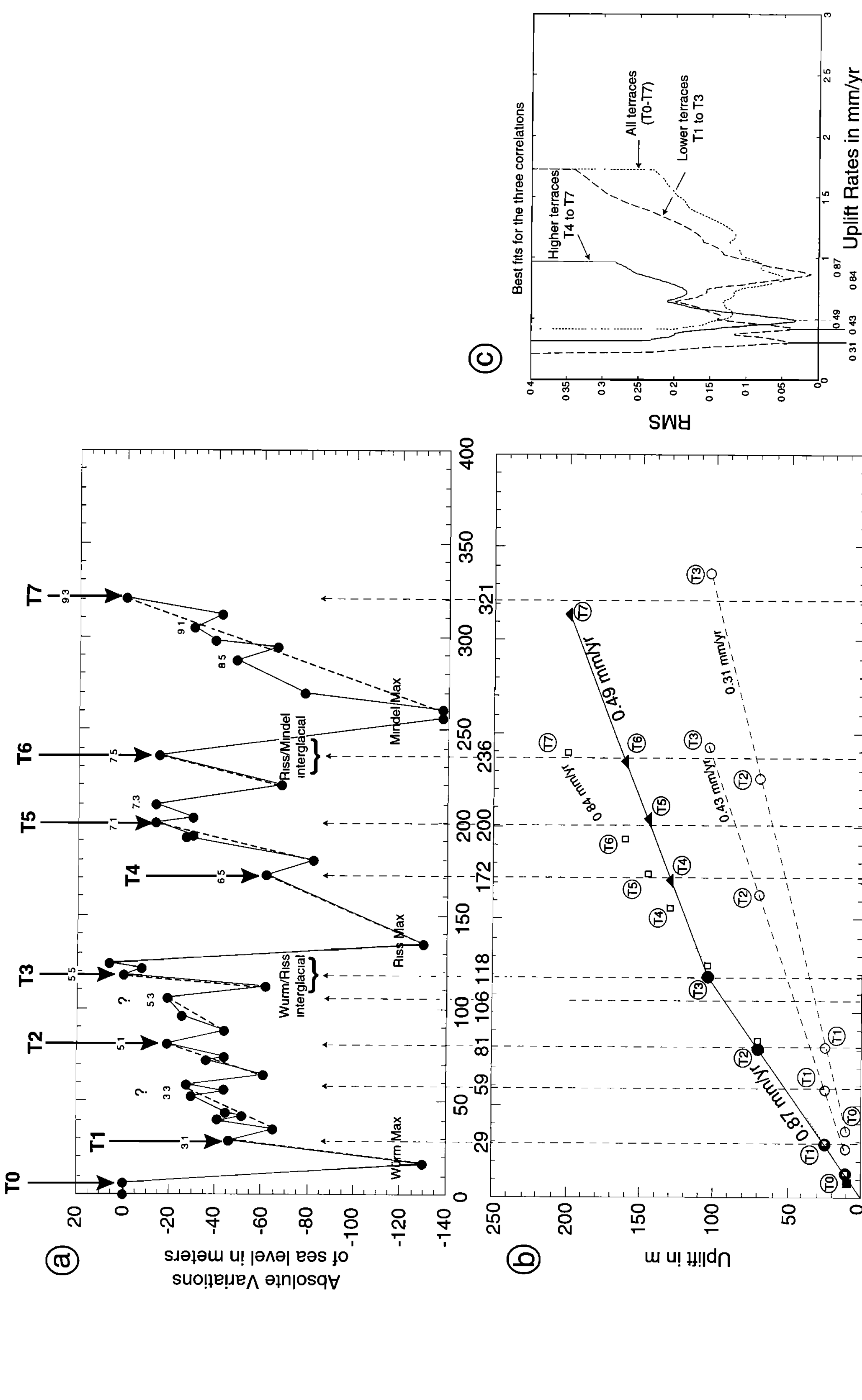
Since the emplacement of T0 appears to be related to rapid post glacial aggradation, we infer similar aggradation, due to climate warming, to have been responsible for the emplacement of all the other, more ancient terraces. The ages of deposition of these terraces would thus correspond to warm interglacials or interstadials prior to the Holocene, and incision of their outer risers would correspond to renewed colder and drier climate conditions, coeval with drops in the level of the sea, located only a few tens of kilometers southwards. In this way, fairly constant and continuous, progressive tectonic uplift would have been punctuated by climate change [e.g. *Avouac et al.,* 1993], with successive, discrete aggradation episodes emplacing new terraces inside older, uplifted ones. According to this interpretation, each terrace riser top ought to be a chron of the end of an interglacial or stadial. In other words, there should be a good correlation between the timing of terrace emplacement and that of sea level highstands, as established, on a global scale, from radiometric dating of marine coral terraces [*Chappell,* 1983; *Chappell and Shackleton,* 1986]. Though the terraces would be regionally correlated to the main interglacial or interstadial aggradational surfaces of the piedmont, locally across the Montello anticline, they were probably strath terraces, as observed in other active thrust environments, particularly in central Asia [*Avouac et al.,* 1993; *Meyer et al.,* 1998].

One additional dating constraint comes from the oldest archeological site known in the Montello region. At Capodimonte, on top of the T'2 flat north of Montebelluna (Plate 2), thousands of arrowheads, flint tools, and other kinds of artefacts testify to permanent human occupation from the beginning of the middle Paleolithic ($\prec$ 37 ka, Musterian industry) till Roman time [*Berti and Boccazzi,* 1956; *Gerharfinger,* 1986; *Capuis et al.,* 1988]. The age of the T2 terrace therefore must predate 37 ka, but not by an implausibly long amount of time.

5. Terrace Uplift Rates and Growth of the Montello Anticline

5.1. Uplift Rates

On the basis of the global sea level curve shown on Figure 7a and using the relative height measurements derived from the profiles in Plate 4 (Table 1), we have tested the existence of a correlation between the ages of


a
Absolute Variations of sea level in meters
T0
T1
T2
T3
T4
T5
T6
T7
Wurm Max
Riss Max
Mindel Max
Wurm/Riss interglacial
Riss/Mindel interglacial
b
Uplift in m
Time in Kyr
0.87 mm/yr
0.49 mm/yr
0.84 mm/yr
0.43 mm/yr
0.31 mm/yr
c
RMS
Uplift Rates in mm/yr
Best fits for the three correlations
Higher terraces T4 to T7
All terraces (T0-T7)
Lower terraces T1 to T3

emplacement of the seven uplifted terraces and sea level highstands. To do this, we make the assumption that the uplift rate remains constant over substantial time spans, though not necessarily during the entire period of terrace deposition.

We calculated ages for uplift rates varying between 0.01 and 3 mm/yr, with increments of 0.01 mm/yr. This provides sets of ages to compare with glacial events dates. All the sea level highstands of the Würm period are taken into account in the correlation. For the Riss and Mindel periods, during which the record is less accurate, we did not attempt to correlate the less prominent highstands (stages 7.3, 8.5, and 9.1).

Figures 7b and 7c shows the results of those different correlations. If only one, uniform uplift rate is taken, only three rates (0.31, 0.43, and 0.84 mm/yr) yield acceptable correlations (RMS $\simeq$ 0.05), and the first two rates only yield correlations for the lowest terraces (T0-T3). With these three rates, however, the age found for T0 is older than 20 ka, and the fits for T2, T3, and T4 at 0.84 mm/yr are poor. In particular, if the 0.84 mm/yr rate is extrapolated back beyond 120 ka, T4 would have been emplaced around 150 ka, during the Riss glacial maximum, which is particularly unlikely. Two different uplift rates, 0.87 mm/yr during the emplacement of the lowest (T0 to T3) and 0.48 mm/yr during emplacement of the highest (T4 to T7) terraces yield the best correlations of all, with RMS smaller than 0.02 and 0.04, respectively. With these two rates, the emplacement of all the terraces corresponds to ages of sea level highstands (Figures 7a and 7b).

This best fit, with a rate prior to 120 ka of about half the most recent rate, suggests that the ages of terraces T1 to T7 range between $\sim$ 30 and $\sim$ 320 ka (Table 1). There are several problems, however. For instance, the calculated age for T6, which is neither the most prominent nor the most continuous terrace, corresponds to one of the most prominent interglacials (stage 7.5, between Riss and Mindel). A possible explanation is that given the relatively small uplift rate and the comparably warm interstadials 7.3 and 7.1, the T5 deposits (stage 7.1) partly merged with those of T6. Two highstands of the Würm period, stages 3.3 and 5.3, find no expression in the terrace sequence. This might be due to the greater tectonic uplift rate. One of the important factors in the formation of fluvial terraces is the power/incision rate ratio of a river. According to most authors [e.g. *Merrits et al.*, 1994; *Bull*, 1991; *Lavé*, 1997] as this ratio increases, so does the likelihood of terrace formation. The incision rate increases with the uplift rate [e.g. *Merrits et al.*, 1994; *Lavé*, 1997]. The power is mainly related to the size of the drainage area [e.g. *Merrits et al.*, 1994]. In the case of the Piave this area has probably not changed much during the last 300 kyr. Consequently, at the end of the Mindel and during the Riss glacials, when the uplift rate was small, so was incision, and the power/incision rate ratio was high, yielding a situation in which broad terraces could form. With the near doubling of the uplift rate, this ratio decreased during the Würm. Thus only small and narrow terraces could form, and some of them could be missing altogether. That a rather abrupt change in the power/incision rate ratio occurred is supported by the shape of profiles transverse to the Biadene paleovalley, which exhibit a gentle average slope for the highest terraces (T4 to T7) and a steeper V-type incision for the lowest ones (T1 to T4) (Plate 3c). The best fit in Figure 7, with an uplift rate increasing by a factor of $\simeq$ 2, from 0.49 $\pm$ 0.1 to 0.87 $\pm$ 0.1 mm/yr, between the emplacement of T3 (121 ka) and T4 (172 ka) is thus in agreement with the geomorphic evidence in Plate 3c, which implies that the Piave only started to cut into the growing Montello anticline after the emplacement of T4.

Such uplift rate values are comparable to, if on the low side of, those found across other growing folds for example in China [e.g. *Avouac et al.*, 1993; *Meyer et al.*, 1998] or California [*Medwedeff*, 1992; *Burbank et al.*, 1996]. Although a present-day uplift rate of $\sim$ 1 mm/yr across the Montello was derived from 33 years of levelling data (1952-1985) by *Balestri et al.* [1994], it is so poorly constrained ($\sigma \geq 1$ mm/yr) that it might be much smaller and cannot be compared at this stage with our mid-Pleistocene-Holocene value.

Changes of terrace uplift rates inferred to be of tectonic origin have been documented in other regions of active plate convergence. In New Zealand a nearly twofold increase in marine terrace uplift rate about 140 ka was attributed to increased convergence between the Pacific and Australian plates [*Bull and Cooper*, 1986]. This was later thought to be an artifact of the less detailed sea level change record prior to $\sim$140 ka [*Pillans*, 1990]. In California, changes in terrace uplift rate across the Ventura Avenue Anticline [*Rockwell et al.*, 1988] have been attributed to changes in the mechanics of folding under a constant shortening rate [*Rockwell et*

Figure 7. a) Curve of absolute variations of sea level versus time, from *Chappell* [1983] and *Chappell and Shackleton* [1986]. Numbers indicate isotopic stages. Arrows represent preferred ages for terraces T0 to T7 derived from correlation in Figure 7b, based on correspondance between fill terrace emplacement and sea level highstand. Isotopic stages 3.3 and 5.3 are not associated with a terrace. (b) Maximum terrace uplift (from Table 1) as a function of calculated ages with uniform uplift rates. (c) Best uplift rates, ranging between 0.31 and 0.87 mm/yr and corresponding to lowest RMS values (normalized RMS misfit of terrace heights for specified ages) obtained in correlation with sea level highstands.

al., 1988]. The modeling presented in section 5.2 also suggests that a change in the geometry of the thrust underlying the Montello anticline can explain the change of uplift rate.

5.2. Modeling the Observed Terrace Profiles

The Montello anticline results from slip on a blind, north dipping ramp and flat thrust system [e.g. *Pieri and Groppi,* 1981; *Slejko et al.,* 1987], (Figure 3). Seismic sections provide constraint on the dip of the underlying ramp (50-40°N [*Pieri and Groppi,* 1981]). The sections also suggest that the depth of the upper tip of the ramp below the surface is between 0.5 and 2 km [*Pieri and Groppi,* 1981].

We model the folding of the terraces following the approach of *King et al.* [1988] and *Stein et al.* [1988], where long-term geological structures are taken to result from the sum of deformations due to successive earthquakes. The seismogenic crust is treated as elastic, with faults represented by cuts. This elastic layer overlies a viscous half-space representing the lower crust. Gravity, erosion, and sedimentation modify the deformation by producing flexure with wavelengths ≥ 10 km. The Montello anticline has an overall width ≤ 6 km. The effects of gravity will therefore be at a scale greater than examined and produce only an apparent baseline shift. We thus ignore such effects and model the fold with dislocations in an elastic half-space. Other authors such as

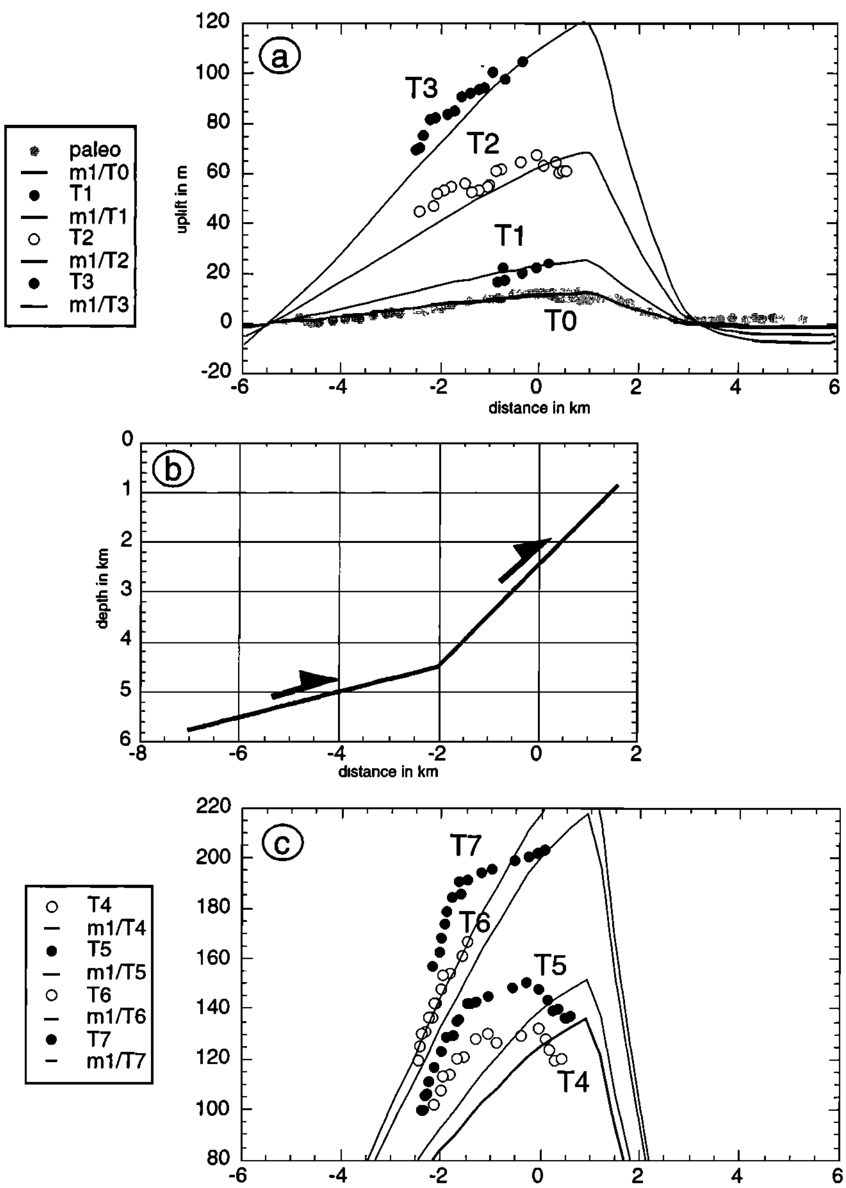


Figure 8. Modeling of terrace folding. (a) Profiles of paleovalley and terrace risers tops (data, circles, and model, lines). (b) Geometry of fault used to model paleovalley (m1). Top of thrust ramp is 0.9 km from surface. The 5-km-long ramp dips 45°N to meet the 5-km-long, 12°N dipping low-angle thrust 4.5-km below surface. (c) Model in Figure 8b provides fit to T4 to T7 profiles.

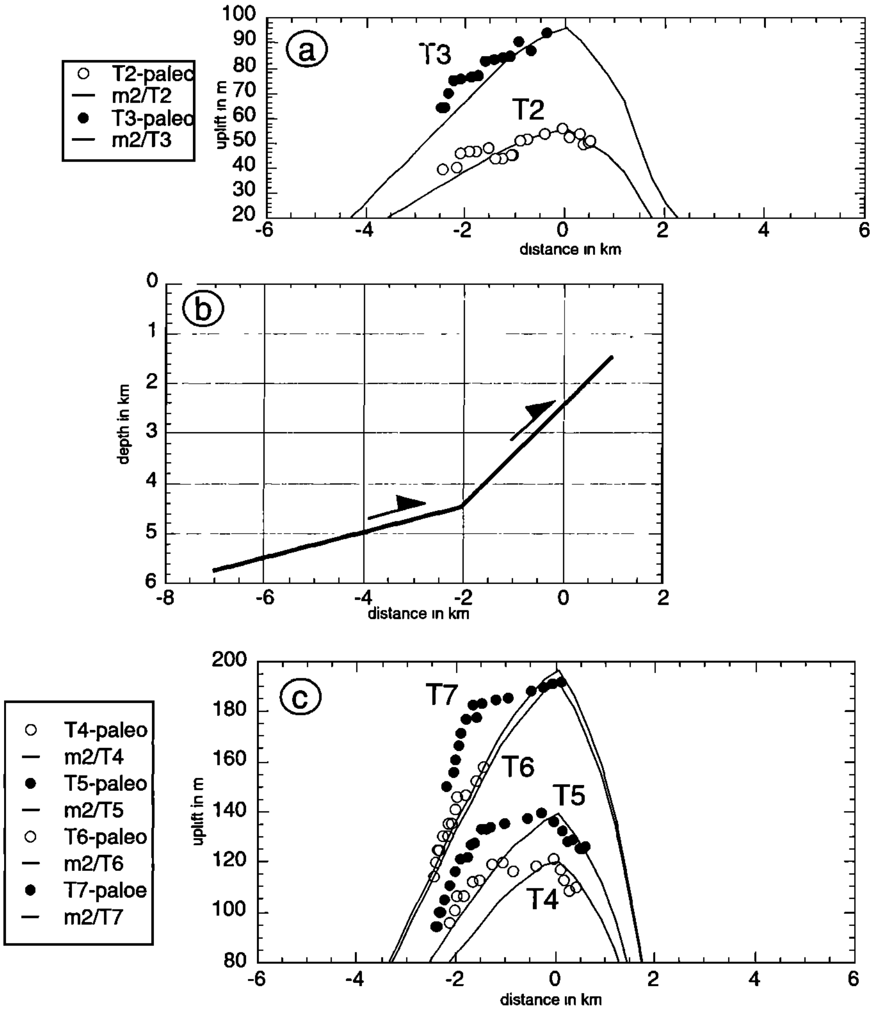


Figure 9. (a) Same as figure 8 with paleovalley shape substracted (data, circles, and model, lines). (b) New geometry of fault used to model T2 and T3 (m2). Top of ramp is 1.5 km from surface. The 4.2-km-long ramp dips 45°N to meet same low angle thrust at same depth as in m1. (c) Fit is still poor with T4 to T7 profiles.

Ward and Valensise [1996] have adopted a similar approach. Since the only parameters we adjust are the slip and geometry of the faults, the model is independent of Youngs modulus and only mildly sensitive to Poissons ratio [*King et al.,* 1988].

A distribution of dislocations can always be found that will reproduce any observed surface deformation. To limit the possibilities, we consider only simple ramp and flat models consistent with the geology and impose constant slip on both.

The depth to the top of the ramp controls both the form of the southern limb of the fold and its position. The position of the base of the ramp, its dip, and the extent of the flat control the form of the central and northern part of the fold. The width of the fold is partly controlled by where, in the model, we choose to terminate the flat to the north. In reality, it must pass into some other structure. We make no attempt to model this since our data would not constrain it.

It is clear that one fault geometry with fault slip increasing with time cannot explain the folding of the terraces because the shapes of the profiles are not self-similar (e.g., Figure 8c). We therefore proceeded as follows. We first found a model explaining the shape of the paleovalley floor T0. The fit is shown in Figure 8a and the model in Figure 8b. The discrepancy with the model in the southern part of the profile is possibly due to sediment accumulation in the Montebelluna fan. While the fits are reasonable for T1, T2, and T3, the apex of the anticline for T2 does not coincide with that for T0 or the model (Figure 8a). Unfortunately, the apex of the anticline is absent for T3.

To find an improved model for these three terraces, we allow for a change of thrust geometry and first subtract the form of the paleovalley from their profiles to remodel them. The adjusted profiles are shown in Figure 9a and the model is shown in Figure 9b. The only change is an increase, by 0.5 km, of the depth to the

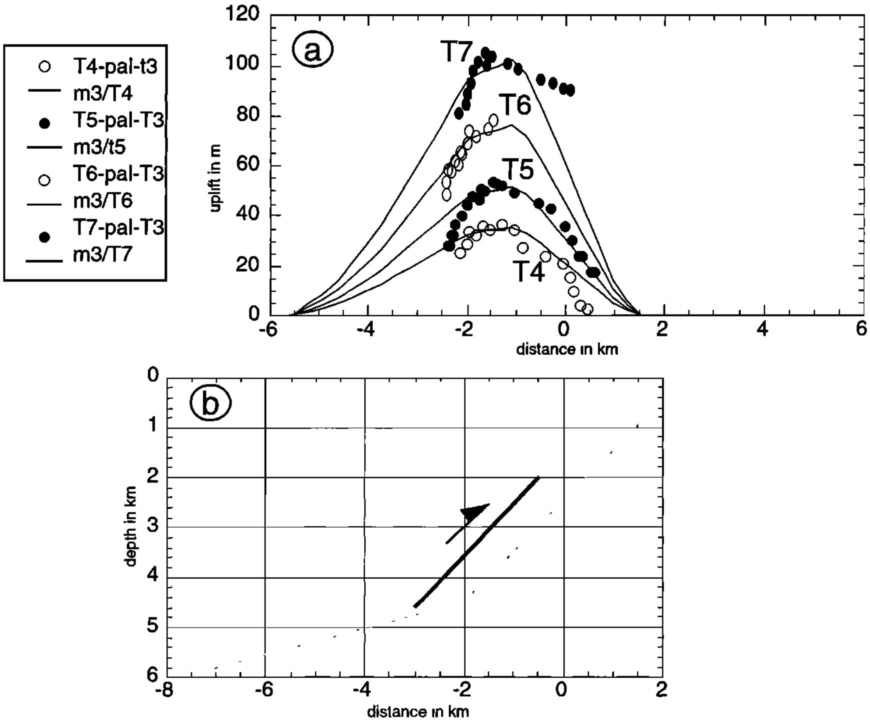


Figure 10. (a) Same as Figure 8 with paleovalley and T3 shape substracted (data, circles, and model, lines). (b) New geometry of fault used to model T4 to T7 (m3). Top of ramp is 2 km below surface. Thrust ramp is 3.5-km-long and dips 45°N.

top of the ramp. The fit to terrace T2 is improved. In Figure 9c, fits to the upper terraces T4-T7 are shown using the same model. It is clear that the modified model that improves the fit to terrace T2 does not improve the fit to the earlier terraces enough. A more substantial change is needed.

The new fits and model are shown in Figures 10a and 10b, respectively. The top of the ramp is dropped a further 0.5 km, and more importantly, the whole ramp is shifted 1 km to the north. No motion on the decollement is required. The fits to the upper terraces are improved except for the southern part of terrace T7. This may be because this terrace shape has been degraded or, more likely, in part misidentified. The slightly concave southfacing slope is not to be expected for a folded surface. The discrepancy with the model at the southern extremity of terrace T4 may be due to lateral action of the paleo-Piave just at the exit of the Biadene cluse.

Although the most straightforward way to model the terraces is to work backwards in time as we have done, it is not the easiest way to visualize how the fold has developed. This is shown in Figure 11. Cross sections in Figures 11a to 11h display the evolution of the fold corresponding to each terrace. Deformation starts with slip on a simple 45° dipping ramp (Figures 11a-11d). We assume that a decollement must have been active previously but find no evidence for substantial movement during the period of emplacement of terraces T7 to T3 (between 321 and 172 ka). After that time the modeling suggests that a shallow 12°N dipping decollement was active and that deformation shifted to a new ramp 1 km ahead of the earlier one (Figures 11e-h). During this evolution the fold axis migrated progressively southeastwards (Figure 11i), as was apparent in the original data set (Plate 4).

The most unsatisfactory outcome of the model is the absence of motion on a decollement linking the Montello to the rangefront between 350 and 150 ka. This may imply that the elastic modeling approach chosen fails to account completely for the early development of the anticline and that another approach would be preferable [e.g. *Rockwell et al.,* 1988].

5.3. Long-Term Folding and Shortening, Slip Rate and Lateral Growth

Together, the results of the modeling and the evidence drawn from seismic sections [*Pieri and Groppi,* 1981], geological exposures [*Dal Piaz et al.,* 1963], and the topography are consistent with a large-scale cross section (Figure 3) that accounts well for present-day shortening and thrusting along the Venetian Alps range front. Although locally, east of Valdobbiadene, the Jurassic and Cretaceous limestones of Mount Cesen are thrust over the Oligocene, there is little morphological evidence of Quaternary movement on this thrust. Instead, much of the present-day slip seems to have been transferred 15 km southeastwards to the Montello ramp by means of a shallow northwest dipping decollement

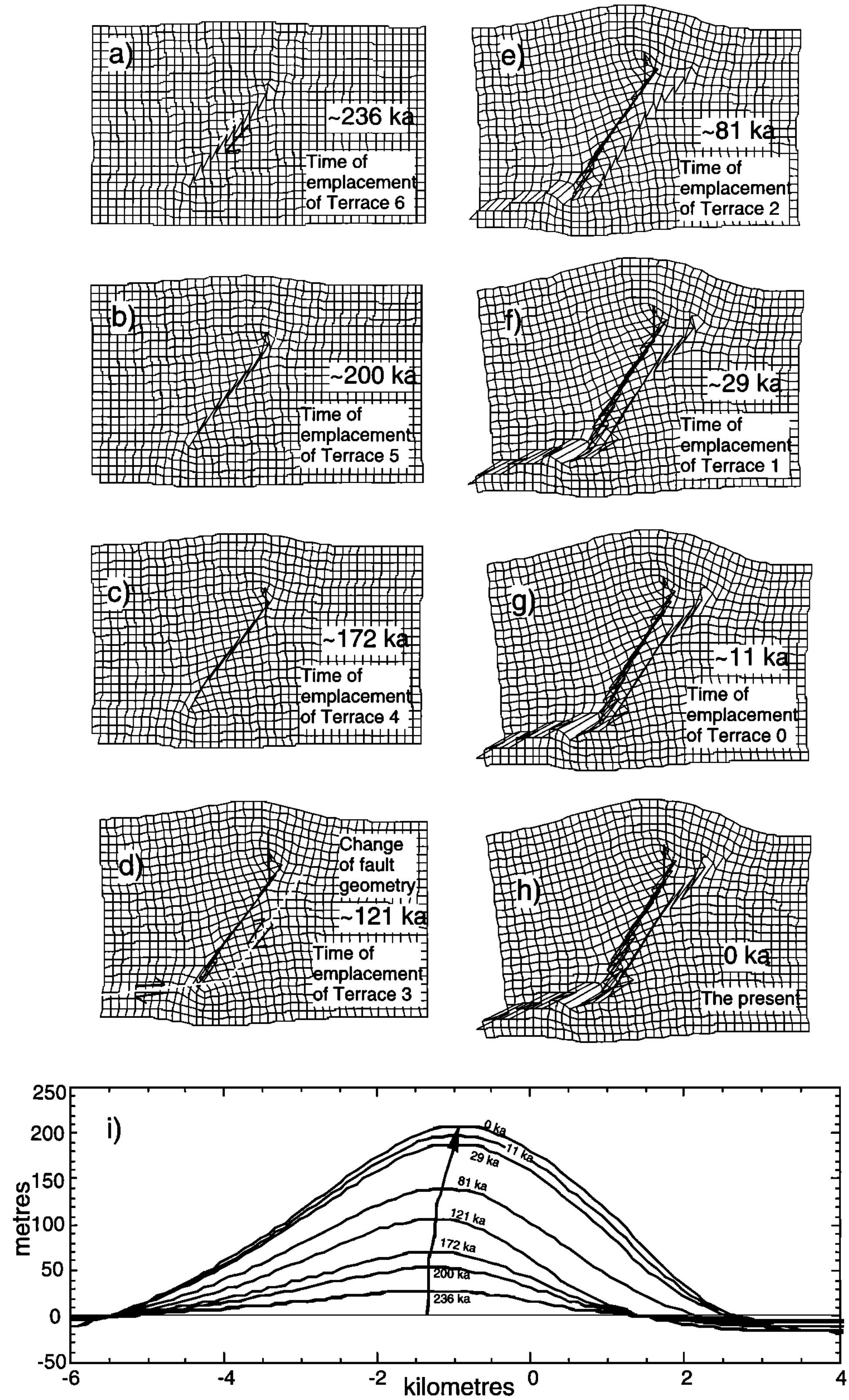


Figure 11. (a)-(h) Evolution of Montello anticline shape in steps corresponding to emplacement of terraces T7 to T0. No vertical exaggeration. Fault slip is 5 times actual slip. (i) Progressive modeled folding from T7 to T0. Vertical exaggeration is 10.

that probably roots into a deeper thrust ramp beneath Mount Cesen. This decollement is probably guided by a rather flat, less competent, sedimentary interface near the base of the Mesozoic cover (Raibl beds, lower Trias), [*Bosellini*, 1989]. Transfer of shortening to the Montello within the foreland must have occurred prior to 350 ka.

The modeling provides estimates of the slip (or shortening) rate on the Montello thrust system, with values ranging between 1.42 and 2.12 mm/yr (Table 1). Given the uncertainties concerning terraces T1 and T7, whose shapes are poorly modeled, we consider the better determined rates to be between 1.77 and 1.81 mm/yr for the youngest terraces and 2.09 and 2.2 mm/yr for the oldest ones. The difference between these slip rates, 15%, however, is not very significant. A constant slip rate of 1.8-2 mm/yr is compatible with offsets that postdate the end of the Pliocene ($\sim$ 2 Ma) on the geological cross section (Figure 3).

The morphological evidence in Plates 3 and 4 indicates that for 300 kyr the Montello anticline kept growing in length as well as in height. A rough measure of the rate of increase in length can be derived from the ages of the terrace riser tops and from the horizontal distance between them at the southwest extremity of the fold. Between the T7 and T4 risers, the along-strike propagation rate of the southwest tip of the anticline was of order of 10 mm/yr (Plate 3c, P1 and P2). Between the T4 riser and the present-day extremity of the fold near Caerano, this rate was of order of 20 mm/yr (Plate 3c, P1 and P2). Thus both the uplift rates and along-strike fold-tip propagation rates increased by a factor of 2 between the emplacements of T3 and T4. Both increases are likely to be due to the same change in fault geometry (from m1, m2, to m3, Figures 8-9-10). While the uplift and propagation rates change, however, their ratio remains constant ($\simeq$ 1/20), which suggests that the anticline is growing self-similarly, as the thrust beneath it. At its northeastern end the Montello fold axis curves and merges with the mountain range. It is thus difficult to determine whether the Montello thrust also grew towards the NE. But the evidence at the SW pericline of the fold is consistent with the observations of other authors [e.g. *Jackson et al.*, 1996; *Meyer et al.*, 1998], which suggests that the length, width, and height of folds reflect the geometry of the thrust ramps that grow and propagate beneath them, maintaining a roughly constant length to depth ratio.

6. Seismic Behaviour of the Montello Thrust and Regional Tectonics

The pattern of deformation observed in the Montello region with a blind thrust system extending into the foreland and rooting under the mountains is very common along active mountain ranges [*Ikeda*, 1983; *Avouac et al.*, 1993; *Gaudemer et al.*, 1995; *Yeats and Lillie*, 1995; *Lavé*, 1997; *Meyer et al.*, 1998]. Near the fronts of large ranges such as the Tien Shan, Qilian Shan, and Himalayas, similar thrust systems have been the source of destructive earthquakes [*Avouac et al.*, 1993; *Gaudemer et al.*, 1995; *Meyer et al.*, 1998].

The historical seismicity in the Montello region is shown in Figure 2. Two series of events, in 1268 and 1857-1860 [*Baratta*, 1901; *Boschi et al.*, 1995; Monachesi and Stucchi, data, 1997], may be related to motion on the Montello thrust (Figure 2). On March 1857, a small shock ($I \sim$VI-VII, $M \sim$4-4.5) hit Valdobbiadene and Pieve di Soligo. On January 1859 a larger shock hit Collalto and the surrounding region, with intensities I = VIII-IX and $M \sim 5.5$. It was followed by numerous aftershocks for the subsequent 4 months. Finally, on July 1860, another shock (I = VIII, $M \sim$ 5-5.5) hit Valdobbiadene. The shape of the isoseismal contours and the location of the epicenters, above the ramping part of the Montello thrust, suggest that this sequence was a cluster of coupled events (as those near Gemona in 1976-1977) that ruptured a significant fraction of this thrust surface. The 1268 ($I \sim$ VIII) earthquake may have done the same, though there is uncertainty on the location and size of this event. Despite even greater uncertainty, one earthquake in 778 ($I \sim$ VIII) may also be related to the Montello thrust ramp. The thrust responsible for ongoing folding of the Montello might therefore have slipped three times in the last 2000 years (778, 1268, and 1859 A.D.). With a slip rate of 1.8-2 mm/yr, an event of magnitude 5.5-6.5 with 1-2 m of slip at depth should occur every 500-1000 years, which is consistent with the time span between the three events.

On the basis of the regional geology and our fieldwork and study of satellite images, we propose, tentatively, that the 1695 Asolo earthquake (I $\sim$ IX) ruptured the range front thrust west of Cornuda, a north stepping feature en echelon with the Montello thrust. In keeping with this general interpretation the seismic events that cluster south and east of Belluno (365 A.D., $I \sim$ VIII; 1873, $I \sim$ X, $M \sim 6.5$; 1893, $I \sim$ VIII, $M \sim$ 5.5-6; 1936, $I \sim$ IX, $M \sim 6$) may have ruptured the northeastward continuation of the Montello thrust. The presence of a lateral ramp oriented roughly parallel to the Lapisino valley is necessary to account for the eastern termination of the Belluno syncline.

If the relative motion of the northern Adriatic plate relative to central Europe were assumed to be roughly the same as that between Africa and the European plate, (8 mm/yr, N15°W) [*De Mets et al.*, 1990; *Anderson et Jackson*, 1987], the 2 mm/yr, N30°W shortening absorbed by the Montello thrust would be only a fraction of the shortening expected across the Alps north of Venice. This would imply either the existence of other blind, ENE-WSW trending thrusts farther south beneath the Venetian plain and within the Adriatic or, more plausibly, that the remaining convergence be absorbed within the Alpine range and along the thrusts that bound it to the north.

Farther northeast along the Venetian range front, the cluster of events in the northern Friuli region reflects motion along thrust faults such as those that generated the 1976 Gemona earthquake sequence [*Amato et al.*, 1976; *Slejko et al.*, 1987] and those mapped from satellite images on Figure 2. By contrast, the large earthquake that occurred in 1511 ($I \sim$ IX-VIII) along the Italian-Slovenian border, despite its uncertain location [*Ribaric*, 1979], might have ruptured one of the active, right-lateral, NNE-SSW striking strike-slip faults whose linear traces cut the northern tip of the Dinaric Alps, north of Trieste (Cividale fault?) (Figure 2). Farther north, the 1348 Villach earthquake (I $\sim$ X) may have been related to strike slip on the Hermagor fault.

In any event, we infer the two most obvious seismic clusters in Figure 2 to be associated with active structures whose geometry and kinematics vary from west to east. A change from roughly NNW shortening on thrusts west of the Tagliamento to predominant dextral motion on NW-SE trending strike-slip faults east of Udine and Trieste would be compatible with the NNW-SSE, Africa-Europe convergence direction in the region. The absence of historical seismicity near Maniago might reflect a seismic gap on the Venetian Alps range front thrust.

7. Conclusion

Our study documents the existence of active blind thrusting, leading to folding of piedmont sediments and to deviation of the course of a large river along the southern front of the eastern Alps. For over 300,000 years the Montello anticline appears to have been growing due to NW directed shortening at a rate of $\simeq$ 2 mm/yr. Such shortening absorbs part of the motion, relative to Europe, of the leading, northern tip of Adria, which is linked to Africa and penetrates into Europe, guided on its northeastern side by NW trending, dextral strike-slip faults [*Benedetti*, 1999]. We infer the entire Venetian Alps range front between Schio and Gemona to be the site of such active thrusting, concurrent with the rather high level of seismicity, one of the highest in the Alps.

At a more detailed level a doubling of the uplift rate (0.5-1 mm/yr) of the anticline apex, coeval with a doubling of the propagation rate of its southwest tip (10-20 mm/yr) probably reflects a change in the geometry of the underlying thrust, between $\sim$ 170 and 120 Ka. Such propagation progressively shifted the course of the Piave by 3 to 4 km before it was captured by the Soligo, at the end of deglaciation about 11 ka. The growth of the Montello anticline, both in height and length, is similar to that found for Wheeler Ridge, in California [*Medwedeff*, 1992; *Burbank et al.*, 1996; *Mueller and Talling*, 1997], for folds in Otago in New Zealand [*Jackson et al.*, 1996], at the foot of the high mountain ranges that rim northeastern Tibet (Qilian, Taxueh, Tanghenan Shan) [*Avouac*, 1991; *Meyer*, 1991; *Avouac et al.*, 1993; *Meyer et al.*, 1998] or in the foothills of the Himalayas [*Talling et al.*, 1995]. All of these authors have pointed out the effect of fold growth on river catchments evolution, but nowhere, to our knowledge, has the effect been reconstructed for a time span of several hundred thousand years and a river as large as the Piave.

Much of our unraveling of the recent deformation of the Montello, an anticline long surmised to have formed in the Pliocene and to exhibit little evidence of Quaternary activity [*Dal Piaz*, 1942], rests on the quantitative study of the fluvial terraces and paleovalley abandoned by the Piave north of Montebelluna. Our conclusions on the deformation and fluvial origin of the terraces confirm those of earlier authors [*Comel*, 1955; *Venzo*, 1977], but our assessment of their ages differs. In order to estimate these ages we have attempted to link all the available evidence resting on ^{14}C radiometric ages with the paleoclimatically orchestrated fluvial and glacial history of the Piave river catchment. Our main conclusion is that the terraces are emplaced as a result of aggradation at times of major global warming following prominent stadial or glacial peaks back to the Mindel period (stage 9.3). Although it is generally not straightforward to correlate the formation of fluvial terraces with climatic change [e.g. *Bull*, 1991; *Merrits et al.*, 1994], the most recent stages of the scenario we propose are not just linked to sea level change but are tied up with the advance and retreat of the Piave Glacier. That warm interstadials or interglacials are preferential times of terrace emplacement at the foot of large mountain ranges is in keeping with recent results obtained in several different regions of central Asia [*Lavé*, 1997; *Van Der Woerd* 1998; *Lasserre et al.*, 1999].

Although the convergence between Africa and Europe is known to induce significant deformation and seismicity north of the western Mediterranean sea, few studies in Europe have so far attempted to assess long-term, ongoing tectonic strain from its quantitative geomorphic signature. Among such studies [e.g. *Meyer et al.*, 1994; *Armijo et al.*, 1996; *Cello et al.*, 1997; *Monaco et al.*, 1997; *Lacassin et al.*, 1998; *Piccardi et al.*, 1999] few have yielded unambiguous results, the most convincing ones being those based on the uplift of marine terraces [e.g. *Armijo et al.*, 1996]. To our knowledge, the Montello is the first clear example of an actively growing fold along either the western, southern, or northern front of the Alps, highest and greatest mountain range between Africa and Europe.

Though this may seem surprising, since features of this kind have long been described in Algeria (seismic growth of a fold was first documented almost 20 years ago at El Asnam [*Philip and Meghraoui*, 1983; *King and Vita-Finzi*, 1981]) California, Asia and elsewhere, it might reflect the fact that given slower plate motion and broader distribution of strain in the western Mediterranean, fewer thrusts absorb enough convergence to leave clear traces at the surface, especially with the additional inconvenience of humid climate and dense vegetation. Alternatively, it might simply reflect the fact that growing folds in the foothills of European ranges have been overlooked, in part due to the widespread belief that the climax of mountain building north of the Mediterranean took place in the Mio-Pliocene and that regions with moderate seismicity do not involve faulting processes comparable to those at work elsewhere. In that case, a thorough search for features comparable to the Montello is urgent if we are to assess properly the seismic hazard of regions surrounding the mountains of Europe.

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