

Formation and Persistence of Extensional Internally Drained Basins: The Case of the Fucino Basin (Central Apennines, Italy)

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- Fucino basin (Central Apennines, Italy) 2
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- **Keywords** 18
- 19 Fucino Basin, Slip-direction, Central Apennines, sedimentary load.
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- 21 **Highlights:**
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- Role of tectonics and drainage systems in the evolution of extensional basins 23 •
- Extensional fault network and sedimentary loading keep the Fucino Basin internally 24 • drained during its evolution. 25
- Sediment load contributes to incrementing of max 30% the faults total geological throw 26 • and locally reorganizes the stress field. 27
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32 Abstract

The interaction between sedimentation/erosion and faulting represents one of the most intriguing topics in landscape and tectonics evolution. Only few studies have been able to document the feedback between faulting and sedimentary loading from field observations. Here, we focus on how sediment loading/unloading influences the dynamics of fault systems in the Fucino Basin, in the Central Apennines (Italy). The Fucino Basin represents a remarkable case study with respect to the other main extensional basins in the Apennines, because of its large dimension, square shape, significant sediment thickness and, its endorheic nature throughout its evolution.

40 We present a detailed structural and geomorphologic analysis of the Fucino Basin and its 41 surroundings, investigating the kinematic and geometry of each main fault strand. The slickenlines analysis reveals multiple families of slip-vectors and timing of activity, suggesting a change in 42 43 extension slip-direction from N240° to N200° during middle Pleistocene. Using a local isostatic model, we estimate that up to the 30% of the vertical geological displacement of the faults, which 44 45 overall ranges from 0.5 to 2.5 km, is related to the sediment loading/unloading. We demonstrate a 46 positive feedback between sedimentation and faulting which may also lead to a re-organization in 47 fault kinematics related to a significant increase in vertical stress. We propose a conceptual model for the permanent endorheic configuration of the Fucino Basin, which includes the effect of 48 49 sediment loading.

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64 **1. Introduction**

65 Extension of the crust generates normal faults and basins where sediments are trapped. Morphologically, extensional basins are generally classified as internally (endorheic) or externally 66 67 (exorheic) drained. In the case of an internally drained basin, sediments are transported and 68 stored within it. If the basin is externally drained, sediments are transported out of the basin and 69 deposited in distal regions. The internally drained stage of extensional basins is generally a 70 transient feature as these basins often evolve into an externally drained basin. Endorheic-exorheic 71 and exorheic-endorheic transitions have been documented in several rifts basins, in compressional 72 and extensional settings, to unravel the feedback between sedimentation-erosion-climate and 73 tectonics (e.g. Berry et al., 2019; Bialas and Buck, 2009; Bucher et al., 2021; Garcia-Castellanos, 74 2007; Garcia-Castellanos et al., 2003; Geurts et al., 2018; Nichols, 2004; Sobel et al., 2003; Strecker 75 et al., 2007). Numerical models of landscape evolution describe the role of strain rate in 76 extensional basin systems, such as Rio Grande and East African rifts (Berry et al., 2019).

Drainage systems and sedimentary budgets of basins impact their fault systems. In fact, the sedimentary budget may influence the fault patterns (e.g. Ballato et al., 2019; Maniatis et al., 2009; Olive et al., 2014; Steer et al., 2014; Turpeinen et al., 2008; Vernant et al., 2013; Whipple, 2009). Local unloading may influence fault slip, and even trigger seismicity (e.g. Calais et al., 2010; Hampel et al., 2009, 2007; Hampel and Hetzel, 2006; Hetzel and Hampel, 2005; Mey et al., 2016). The feedback between faulting and sedimentary loading may thus be particularly relevant for internally drained basins, where the sediments are expected to be more abundant.

84 In this paper, we investigate the relation between faulting and sedimentation, focusing on the 85 Fucino Basin in the Central Apennines (Italy, Fig. 1). Here, normal faulting and erosion processes 86 shaped the landscape, producing a basin-and-range-type mountain belt forming both endorheic 87 and exorheic basins. The main tectonic basins are Rieti, Leonessa, L'Aquila, Sulmona, Fucino and 88 Subequana Valley (Fig. 1a). Some of these basins are externally drained by the Aterno-Pescara, the 89 Sangro, the Salto-Tevere and the Liri rivers. Other basins are internally drained such as the Fucino, 90 the Campo Felice, the Piani di Pezza and the Altopiano delle Rocche basins (Fig. 1b). Despite the 91 fact that all these basins evolved at the same time and under the same tectonic regime (e.g. 92 Cosentino et al., 2017), they show remarkable differences in terms of dimensions and sediments 93 thickness. We focus in this study on the Fucino Basin, because of: (1) its large dimensions (~200 km²); (2) an almost square shape due to the presence of two orthogonal faults (Fig.1) rather than 94 95 an elongated shape (e.g. Sulmona or L'Aquila Basin; Fig.1); (3) sediment thickness which reaches almost 1 km (Cavinato et al., 2002) compared to 350 m (e.g. Sumona Basin, Subequana Valley,
L'Aquila; Gori et al., 2017; Nocentini et al., 2018, 2017), and 600 m (eastern L'Aquila basin;
Nocentini et al., 2018). The Fucino Basin has remained endorheic whereas the other intermontane
basins underwent through endorheic-exorheic transitions and vice versa (e.g. D'Agostino et al.,
2001b; Geurts et al., 2019, 2018).

101 The aim is to understand the conditions that allowed the Fucino Basin to preserve its internal 102 drainage during its entire evolution. We perform a detailed structural analysis to define the 103 geometry and the kinematic evolution of the main faults around the Fucino Basin. We also provide 104 a geomorphological analysis including swath profiles, steepness index, local relief and slope maps, 105 and drainage divide stability analysis (Forte and Whipple, 2018) to investigate the linkage between 106 erosion and faulting. Our results reveal that sedimentary processes increase the fault offsets 107 around the Fucino area up to 30%, mainly due to both hanging wall loading and footwall 108 unloading. We interpret the observed change in strain direction as an increase of the vertical 109 stress due to the sedimentary loading. We speculate that the coexistence of two orthogonal faults 110 (which is the major difference between the Fucino Basin and any other intermontane depression 111 in the Central Apennines) combined with the sediment (un)loading is responsible for enhancing 112 the subsidence and the stress reorganization. This prevents the Fucino Basin to be captured by 113 nearby rivers, favoring the long-lasting endorheic state of the basin.

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115 **2. Geological Setting**

2.1. Structural setting, kinematic and throw distribution along the Central Apennines

117 The Apennine mountain belt is an asymmetric Neogene fold-and-thrust accretionary wedge 118 produced by the subduction of the Adria Plate (Cipollari and Cosentino, 1996; Cosentino et al., 119 2010; Merlini and Mostardini, 1986; Patacca et al., 1992). Crustal accretion is progressively 120 younger eastward toward the foreland and is followed by extensional deformation. Extensional 121 deformation produced a set of normal faults striking sub-parallel to the older thrusts and trending 122 mainly NW-SE. The most active normal fault system is located on the top of the wedge and it is 123 responsible for the formation of intermontane basins. The only exception to this pattern is the 124 Avezzano-Buzzi fault system which strikes instead NE-SW from the Fucino to the Sulmona basin 125 (Fig. 1a; Gori et al., 2017).

Several studies documented and explored the geometry and kinematics of the normal faults,which are particularly well exposed on the carbonate deposits of the Central Apennines (Cowie et

128 al., 2017; Cowie and Roberts, 2001; Faure Walker et al., 2012, 2010, 2009; Galadini, 1999; Galadini 129 et al., 1997; Galadini and Galli, 2000; Gori et al., 2017; Papanikolaou et al., 2005; Piccardi et al., 1999; Pizzi and Pugliese, 2004; Roberts et al., 2004; Roberts and Michetti, 2004; Salvi et al., 2003; 130 131 Smeraglia et al., 2016; Wilkinson et al., 2015); investigating also the role of tectonics in landscape 132 evolution of the Central Apennines (e.g. Bubeck et al., 2015; Geurts et al., 2019, 2018; Whittaker et al., 2007). Structural analysis of the faults shows a prevalent NNE-SSW to ENE-WSW direction of 133 134 extension (Faure Walker et al., 2010), with a mean direction oriented NE-SW. This is consistent 135 with focal mechanisms, borehole breakout data, inversion of micro-earthquakes and geodetic 136 strain (Fig. 1a; D'Agostino et al., 2001a; Montone et al., 2012, 1999). In several locations 137 surrounding the Fucino basin, multiple direction of slickenlines have been measured and: (1) 138 averaged and interpreted as due to fault growth and linkage (e.g. Roberts and Michetti, 2004), (2) 139 associated to block rotation (Piccardi et al., 1999) or in general to a previous tectonic deformation 140 (Galadini, 1999). The geological throw of those normal faults ranges from 500 to 2500 m and is 141 larger on the Sulmona, L'Aquila and Rieti faults (Morewood and Roberts, 2000; Roberts and 142 Michetti, 2004). A maximum cumulative throw of 6600 \pm 800 m has been estimated along a 143 section crossing the Fucino and the Sulmona basins (Roberts and Michetti, 2004).

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2.2. Structural setting of the Fucino Basin

146 The Fucino Basin is one of the largest extensional basins in the Central Apennines and is at an 147 elevation of 600 m a.s.l. It is bordered to the east by the San Benedetto, Pescina and Collarmele 148 faults, which extend southward and northward to become the Gioia de Marsi and the Ovindoli-149 Celano faults, respectively; by the Vallelonga fault on the western side (Fig. 2) and bordered to the 150 north by the Tremonti and Velino-Magnola faults. These two sets of orthogonal normal faults give 151 the Fucino Basin an almost square shape. Other extensional basins are more elongated and 152 controlled by only one faults strand (Fig.1a). The existence of the two sets of orthogonal faults as 153 well for the age of the Avezzano-Buzzi faults is unclear (Tremonti and Celano-Aielli; Fig. 1). It has 154 been suggested that the Avezzano-Buzzi fault was active during the Apennines compressional 155 phase as a lateral ramp and then has been reactivated during the Plio-Quaternary extension (e.g. 156 Gori et al., 2017). Further north, two other main faults are present: the Piani di Pezza and Campo 157 Felice (Figs. 2 and 3).

158 The Fucino Basin is surrounded by four major exorheic catchments: the Salto-Tevere and Liri 159 basins to the west and the Sangro and the Aterno-Pescara basins to the east (Fig. 1b). On its northern side, the Fucino Basin is delimited by small endorheic catchments, such as the Campo
Felice, the Piani di Pezza and the Altopiano delle Rocche basins (Fig. 1b).

Paleoseismological trenching across these structures show evidence of several large earthquakes in the last 7 ky affecting the Piani di Pezza-Ovindoli, Gioia de Marsi and Vallelonga faults with a slip from submeter to meters (D'Addezio et al., 1996; Galadini and Galli, 1999; Michetti et al., 1996; Pantosti et al., 1996; Salvi et al., 2003). Slip rate deduced by ³⁶Cl cosmogenic dating along the faults escarpments ranges from 0.2 to 1.2 mm/yr (Benedetti et al., 2013; Palumbo et al., 2004; Schlagenhauf et al., 2011). Cosmogenic analysis and paleoseismological trenching attest an increase of Holocene slip around the Fucino Basin (Benedetti et al., 2013).

169 Interpretation of seismic lines and borehole stratigraphy data provide depth, geometry, and 170 maximum sediment thickness of the Fucino Basin (Cavinato et al., 2002). The sedimentary infill is 171 up to 950 meters thick, and consists of lacustrine, alluvial and fluvial sediments deposited from late Pliocene onward (Cavinato et al., 2002; Mondati et al., 2021). Two sedimentary sequences 172 173 (sequence 3 and 4, Fig. 2; Cavinato et al., 2002), separated by an unconformity, indicate the 174 occurrence of main extensional or climate events during the late Pliocene and since middle 175 Pleistocene (Cavinato et al., 2002). The Piani di Pezza and Campo Felice faults are younger than 176 the Fucino faults, and have been active since the middle Pleistocene (Giraudi, 2012; Giraudi et al., 177 2011; Villani et al., 2015).

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179 **3. Methods**

180 **3.1.** Structural analysis

181 We carried out a structural analysis around the Fucino Basin and measured the strike and dip of 182 the main faults and the pitch angle or the azimuth of the slickenlines. The sense of shear of these 183 structures has been deduced by kinematic indicators such as Riedel shear and lunate structures. In 184 case of multiple slickenlines, cross cutting relationships provide the relative chronology. We 185 plotted the analyze data using the 'Daisy' software (Salvini, 2004). Figure 3 shows the transport 186 direction of the hanging wall with respect to the footwall of each measurement site and the 187 cumulative along each fault (stereoplot; Fig. 3). Field observations and frequency analysis of the 188 azimuth of the slip-directions were performed to distinguish between different slickenside populations. In supporting information (Figs. S3, S4 and S5), we show data and gaussian 189 190 distribution of the slip directions for each fault strand. This analysis, together with the cross191 cutting observations, is essential to distinguish between a dispersion of data collection along a 192 mean direction and/or two distinct clusters.

193 Serial cross sections provide the throws along faults using geological markers and corrected for the 194 apparent angle if sections were not orthogonal to the fault strands (Figs. S3, S4 and S5 supporting 195 information). Since multiple fault slip-directions are present, the fault throw cannot be estimated 196 easily, thus we estimate the vertical component of the fault throw.

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- 198 3.2.

Morphometric quantitative analysis

199 We perform a morphometric quantitative analysis, including swath profiles, maps of local relief 200 and slope, river network analysis (k_{sn}) , and drainage divide stability analysis to investigate the landscape and the drainage area of the Fucino Basin. All these analyses have been realized by 201 202 using SRTM DEM topographic data (~90 in horizontal raw m resolution) 203 (http://srtm.csi.cgiar.org/srtmdata/) in GIS environments and MatLab. Two swath profiles, 204 respectively NW-SE and SW-NE trending, are extracted from SRTM DEM data by using 205 TopoToolBox (Schwanghart and Kuhn, 2010). The elevation along each profile is sampled each 100 206 m into an observation window 5 km wide. The final plot represents the patterns of maximum, 207 minimum and mean topography (Isacks, 1992).

208 We perform a divide stability analysis around the Fucino Basin following Forte and Whipple (2018). 209 This analysis provides a snapshot of a possible watershed migration by a comparison of four 210 metrics across divide: gradient, local relief, χ (a proxy of steady-state channel elevation; see Willet 211 et al., 2014), and elevation. We do not discuss the χ outcome since it is strongly influenced by the 212 choice of the outlet elevation (Whipple et al., 2017) and in this respect, the Fucino base level 213 elevation is too different from any other reasonably chosen outlet for the external rivers (e.g. sea level, change from bedrock to alluvial channel). In the analysis we use 10⁴ m² as reference area 214 215 and 1000 m as radius of the local relief. We chose 10 segments along the main drainage divide 216 from different location the Fucino Basin. Using free MatLab in а tool 217 (http://github.com/amforte/DivideTools; Forte and Whipple, 2018), we plotted the relative results 218 in a single graph (Fig. 8b) where the error bar is the standard deviation derived from the 219 histograms included in the supporting information (Fig. S17). This plot provides informations of 220 the stability or of the direction of migration of each divide segment (Fig. 8b).

221 The slope and the local relief maps are created in a GIS environment. The latter represents the 222 result of the subtraction of maximum and minimum topography elevation within a circular moving window of 5 km in radius. This proxy allows in the first order to identify (assuming the same lithology, as for the case for this region) incised sectors, and in comparison with faults, their bearing with tectonic.

226 The map of the normalized along-channel variation in steepness index (k_{sn}) has been elaborated by 227 TopotoolBox (Schwanghart and Kuhn, 2010). K_s is a morphometric parameter which, together with 228 the concavity index (ϑ), describes the shape of a river longitudinal profiles providing information 229 about the state of equilibrium of the drainage systems (Duvall et al., 2003; Kirby et al., 2003; Kirby 230 and Whipple, 2012; Snyder et al., 2000; Zaprowski et al., 2005). Usually, to compare river profiles 231 irrespective of basin drainage area, the k_s is normalized (k_{sn}) by using a reference concavity index 232 value in a range from 0.3 to 0.8. The map of the along-channel variation in steepness index shows 233 (Fig. 8e) the changes in channel slope allowing an easy identification of knickpoints and their 234 possible correspondence with tectonic structures or rock type changes.

Although these analyses are strongly affected by the erodibility of the rock types, the study area (Fig. 2) is composed mainly by one lithology: the Meso-Cenozoic limestone (Fig. 2). Thus, in the study area, any variation in incision rates can be associated to climate or tectonic events rather than lithological variations.

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240 **4. Results**

241 4.1. Structural data

242 The Fucino Basin has a square shape, controlled mainly by Pescina-San Benedetto-Collarmele set 243 of faults trending NW-SE to the east, and the Tremonti and Velino-Magnola faults trending WSW-244 ENE and NW-SE, to the north (Figs. 2 and 3). We integrated our structural data and geological 245 displacements (long-term displacement; see Puliti et al. (2020) for details) for each fault strand are 246 compiled with data from previous studies concerning the escarpments height (coseismic rupture 247 providing the short-term displacement; see Puliti et al. (2020) for details), the morphologic 248 displacements (morphologic evidence of faulting such as triangular facets providing the mid-term 249 displacement; see Puliti et al. (2020) for details), the age of the sediments and the slip-rate from 250 cosmogenic dating and paleoseismological trenching.

All the structural data collected in the field are listed in the 'Supporting information' (Table S1 andFig. S1).

• The Campo Felice fault (CF) borders the endorheic basin to the NE. The fault strikes for ~12 km with an average direction of N125° (Figs. 2 and 3). We measured two senses of motions: one

oriented N202° and one, measured only at the eastern tip of the fault, oriented N263° (Table S1 and Fig. S3; supporting information). Using the Cretaceous bauxite levels as marker we estimated a maximum vertical geological offset of ~1500 m (Fig. S3 supporting information). The maximum morphologic displacement along the CF fault is 230 m (Benedetti et al., 2013), the height of the most recent escarpment is ~10 m (Wilkinson et al., 2015) and the Holocene slip rate is ~1 mm/y (Benedetti et al., 2013). Although the onset of the CF is unconstrained, the oldest deposits within the basin are middle Pleistocene in age (Giraudi, 2012; Giraudi et al., 2011).

• The Ovindoli-Celano fault (O) strikes for ~12 km with a N-S orientation controlled at depth by the
 Ovindoli-Monti D'Ocre thrust front (De Vittorio and Faccenna, 1987). The structural analysis shows
 two senses of motion oriented N230° and N183° but no crosscutting relationship have been
 observed (Fig. 3 and Fig. S4 within supporting information).

266 • The Piani di Pezza fault (PP) represents the northern prosecution of the Ovindoli-Celano fault 267 and borders the endorheic basin to the NE (Figs. 2 and 3). The fault strikes for ~5 km and it is 268 oriented from WNW-ESE in the western region, to NW-SE in the eastern region. Two sets of 269 striations (PP2 in Figs. 2 and 4C) with clear cross cutting relationship indicate two directions of slip 270 oriented N224° (older event) and oriented N167° (younger event) (Table S1 and Fig. S3; Supporting 271 information). We estimated a maximum vertical geological offset of ~1800 m (Fig. S3 Supporting 272 information). The height of the most recent escarpment is ~5.7 m (Villani et al., 2015). Trench 273 exploration gives a Holocene slip rate ranging between 0.6 to 2.3 mm/y (Pantosti et al., 1996). 274 Base on the oldest deposit within the basin the onset of the PP fault is middle-late Pleistocene 275 (Cinti et al., 1992; Giraudi and Frezzotti, 1997; Pantosti et al., 1996; Villani et al., 2015).

276 • The Velino fault (V) borders the SW flank of the Velino Mountain (Fig. 2). It strikes for ~14 km 277 and it is oriented N150° (Fig. 3). The Velino fault structural analysis shows one set of striations 278 oriented N200° (Fig. S2; Supporting information). We estimated a vertical geological offset of 279 \sim 1000 m (Fig. S3; Supporting information). The maximum morphological displacement along the 280 Velino faults is ~750 m (Schlagenhauf et al., 2011) while the average scarp height is ~9.5 m 281 (Benedetti et al., 2013). Even though the onset of extension for this structure is not well 282 constrained, the Velino fault is considered active since the middle-late Pliocene (Cavinato et al., 283 2002).

The Magnola fault is the eastern portion of the Velino fault and borders the southern flank of the
 Magnola Mountain striking WNW-ESE (Figs. 2 and 3). Slip-directions cluster around N199° (Table
 S1 and Fig. S4; supporting information). At the measurement site M2 (Figs. 3 and 5), two families

of striations and cross cutting relationships indicate two directions: one recent oriented N195° and 287 one older oriented N235° (Fig. 5, M2). Using the stratigraphic contact between Miocene and 288 289 Cretaceous rocks we estimated a vertical geological offset of ~2500 m (Fig. S4; supporting 290 information). The morphologic displacement along the Magnola faults is up to 800 m 291 (Schlagenhauf et al., 2011) while the escarpment height is up to 20 m with a slip rate of ~1.3 mm/y 292 (Benedetti et al., 2013; Galli et al., 2012). The Magnola and Velino faults are consider active since 293 the middle-late Pliocene (Cavinato et al., 2002), but the exact onset of extension is not 294 constrained.

295 • The Tremonti fault borders the homonymous mountain, and it represents the direct north-west 296 rim of the Fucino Basin (Figs. 2 and 3). It regionally strikes WSW-ENE. The slickenlines analysis shows multiple striations oriented N210°, N126° and N266° (Table S1 and Fig. S4; supporting 297 298 information). At site TRM4, we observed striations oriented N183° crosscutting the ones oriented 299 N261° (Fig. 6). The vertical geological displacement is ~1000 m (after Cavinato et al, 2002), 300 morphologic displacement is ~450 m and the escarpment height is ~3.5 m with a slip rate of ~0.2 301 mm/y (Benedetti et al., 2013). The onset of faulting is debated due to its possible activation during 302 the Apennines orogenesis as a lateral ramp (Gori et al., 2017). Cavinato et al. (2002) proposed a 303 late Pliocene onset of activation.

The Celano-Aielli fault represents the eastern portion of the Tremonti fault, located east of the
 San Benedetto fault (Fig. 2). It strikes NE-SW, and we measured three main senses of motion: one
 N267°, one N209° and one N98° (Fig. 3 and S5 supporting information). The Celano-Aielli fault is
 considered to be active from middle-late Pliocene (Cavinato et al., 2002; Gori et al., 2017).

The Aschi Alto (or San Sebastiano) fault represents the southern portion of the Pescina Fault (Fig.
2). It strikes NW-SE and show two sets of striations; N214° and N162° (Table S1 and Fig. S5;
supporting information). The vertical displacement is ~500 m (Praturlon, 1968) (Fig. 2), the
escarpment height is ~4 m, and the slip rate is ~0.5 mm/y (Benedetti et al., 2013). The Aschi Alto
fault is active from middle-late Pleistocene (Cavinato et al., 2002; Gori et al., 2017).

The Gioia Dei Marsi fault represents the southern portion of the San Benedetto fault (Figs. 2 and
and strikes N120°. We measured two senses of motion with cross cutting relationships (Fig. 3
and 7). The older striation is oriented N182° while the younger striation is oriented N225° (Fig. 7,
Table S1 and Fig. S5; supporting information). The vertical geological displacement is of ~400 m
(Praturlon, 1968) (Fig. 2). The Gioia Dei Marsi fault is considered active since middle-late Pliocene
(Cavinato et al., 2002; Gori et al., 2017).

The Vallelonga (or Trasacco) fault borders the valley located in the southern rim of the Fucino
Basin (Figs. 2 and 3). It strikes ~20 km with a NW-SE orientation. We measured two slip-directions
and a cross cutting relationship: one younger oriented N253° and one older oriented N171° (Table
S1 and Fig. S5; supporting information). The vertical geological displacement is ~300 m (Praturlon,
1968), the maximum morphologic displacement is ~250 m and the maximum escarpment height is
~4.7 m with a slip rate of ~0.2 mm/y (Benedetti et al., 2013).

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326 **4.2.** Geomorphological and morphometric analysis

The swath profile A-B (Fig. 8a) shows a marked difference in the general topographic setting between the northwestern (at the border between the Fucino and the Salto R. basin) and the southeastern side (at the border between the Fucino and the Sangro R. basins). The northwestern side shows a maximum topography lower (< 1000 m) than the southeastern side (~2000 m) and a minimum elevation only ~50 m lower than the Fucino Basin one. To the SE, the minimum topography is above 1000 m of elevation, with a peak, representative of the drainage divide, around 1300 m.

The swath profile C-D runs orthogonal to the main tectonic structures of the area, highlighting the influence of tectonics on landscape (Fig. 8a). In general, the pattern of minimum, mean, and maximum elevation is similar. To the SW, the Liri R. valley interrupts the topography decreasing towards the Fucino Basin plain. Eastward, beyond the plain, the topography increases regularly up to ~1500 m and then decreases towards the Aterno R. valley.

Our analysis of the drainage divide stability (Fig. 8b; and Fig. S10 Supporting information) suggest overall stability in the Fucino Basin. The maps of local relief, slope and k_{sn} (Figs. 8 c, d, e) show a similar pattern with the highest values (local relief of 1200-2100 m, k_{sn} of 80-500 and slope >30°) corresponding with the footwalls of normal faults and hanging walls of thrusts. The remainder of the area is characterized by medium-low values with the lowest ones (local relief and k_{sn} ranges of 0-300 m and 0-20, respectively) associated with the bottom of main fluvial trunks (Figs. 8 c, d, s).

345

346 **5.** Discussion

5.1. Synthesis of the structural data and comparison with previous studies

Previous studies analyzed the structural setting of the region (Faure Walker et al., 2012, 2010,
2009; Galadini, 1999; Galadini et al., 1997; Galadini and Galli, 2000, 1999; Morewood and Roberts,
2000; Papanikolaou et al., 2005; Piccardi et al., 1999; Pizzi and Pugliese, 2004; Roberts and

351 Michetti, 2004; Salvi et al., 2003; Smeraglia et al., 2016; Wilkinson et al., 2015). Our field 352 measurements, and the differences with previous studies are illustrated in the supporting information. Nine figures and one table, synthesized field data measurements (Figs. S7 to S15 and 353 354 Table S2; Supporting information; Morewood and Roberts, 2000; Piccardi et al., 1999; Roberts and Michetti, 2004; Smeraglia et al., 2016; Faure Walker et al., 2009; Wilkinson et al., 2015). 355 356 Measurements from this and previous studies are similar, in some case identical (Fig. S7 to S15; 357 Supporting information). Stereoplots from all previous works show the presence of multiple slip-358 directions that have been interpreted: (1) as due to block rotation (as Piccardi et al., 1999), (2) 359 older tectonic phase (e.g. Galadini, 1999); or (3) have been averaged on a single direction 360 (Morewood and Roberts, 2000; Roberts and Michetti, 2004) to demonstrate variation of slip-361 direction along-strike (Roberts and Michetti, 2004). In this work, the crosscutting relationship from 362 field observations (Figs. 4C, 5C, 6C and 7B), and frequency distribution of population of slip-363 directions (rather than average), has been used as criteria to distinguish between one or two sets 364 of striations. At several sites, we define two distinct sets of slickensides (e.g. sites PP2, M2 and 365 TRM4), that we attribute to distinct episodes of transport directions along the faults. This 366 approach and result are in agreement with earlier study along the Tremonti fault (Galadini, 1999 and Piccardi et al., 1999; Smeraglia et al., 2016). 367

368 Our structural analysis reveals that the faults around the Fucino Basin show normal to transtensive kinematics, and are oriented from NW-SE to NNW-SSE, with the only exception of the Tremonti-369 370 Celano-Aielli fault (western tip of the Avezzano-Buzzi system; Fig. 1) which strikes NE-SW. Two main slip-directions can be distinguished: N230° to N260° and N200° to N220°. Few faults also 371 372 exhibit slip-direction of N120° to N170°. In the northern region, slickenlines crosscutting 373 relationships indicate a coherent pattern of slip-direction change from N240° to N200° 374 (constrained with the M2, PP2 and TRM4 measurement sites, Figs. 4, 5, 6 and Fig. S6; supporting 375 information), in good agreement with previous works (e.g. Piccardi et al., 1999; Smeraglia et al., 376 2016). Conversely, in the southern regions, slip-directions change from N170° to N220° 377 (constrained with the GdM3 measurement site; Fig. 7 and Fig. S6 supporting information).

The vertical displacement on the faults increases, as expected, towards to the center of the Fucino Basin. The southern faults, including the Vallelonga, Aschi Alto and Gioia de Marsi, show an offset of ~500 m. The northern faults, including Campo Felice and Piani Di Pezza faults, have larger offsets ranging between 1000 and 1800 m, except for the Magnola fault, which displays an exceptional vertical throw of ~2500 m. Offset has been estimated by direct field sections and updated 1:50,000 geological maps excluding for Vallelonga, Gioia dei Marsi and Aschi Alto where
we used the updated "1:100.000 Foglio Geologico Sora N152" (Praturlon, 1968). Some of our
estimates are in accord with previous ones (e.g. Magnola fault; Morewood and Roberts, 2000).
Others instead differ from previous ones, as for the case of Vallelonga where we estimate ~300 m
(Praturlon, 1968) while others, in the same section, estimated up to ~1400 m (Roberts and
Michetti, 2004); or the case of Piani di Pezza fault where we estimate ~1800 m of offset, while
Roberts and Michetti (2004) estimated only 400 m (See Fig. S16 in the Supporting Information).

The onset of extension in the study area is not contemporaneous on each fault strand. The Tremonti, San Benedetto and Magnola-Velino faults are the oldest structures (Cavinato et al., 2002; Gori et al., 2017), formed probably during the late Pliocene, while faults located on the southern and northern side of the basin formed since the early-middle Pleistocene (Cavinato et al., 2002; Cinti et al., 1992; Giraudi et al., 2011; Giraudi and Frezzotti, 1997; Villani et al., 2015).

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5.2. Morphometric observations and drainage divide position of the Fucino Basin

397 The landscape of the area is characterized by deep valleys and steep ranges. The highest values of 398 local relief, slope, and k_{sn} correspond to both extensional and compressive structures, suggesting 399 that tectonics influences the landscape. Evidence of this tectonic control is also shown by the 400 current positions of the drainage divides that are mainly located on the topographic culminations 401 related to thrusts anticline and normal faults, especially on the eastern-north/eastern and on the 402 southwest sides (Fig. 8). Moreover, low values of local relief, slope, and ksn at in summit areas of 403 mountains suggest that these sectors have not been yet reached by the erosion wave induced by 404 the regional uplift likely started around 3 Ma (e.g. Cosentino et al., 2017, D'Agostino et al., 2001b; 405 San Jose et al., 2020).

406 In the Central Apennines, D'Agostino et al. (2001b) proposed that although the Fucino Basin has 407 been internally drained throughout its evolution, the original size has been reduced by the Salto 408 River capturing (Figs. 1 and 8a). Here, the divide is located in an almost flat area (see slope and ksn 409 in Fig. 8d and e) filled by late-Pliocene/early-Pleistocene lacustrine sediments (Centamore, 2004) 410 and by the fluvio-glacial Majelama Valley fan (Frezzotti and Giraudi, 1992). The low energy 411 lacustrine environment is inconsistent with the current base level of the Salto R. (Tyrrhenian Sea), 412 suggesting a southeastward drainage divide migration. The Salto-Fucino divide represents the potential outlet for the Fucino Basin, but the ~30 ka Majelama Fan emplacement (Fig. 2; Frezzotti 413 414 and Giraudi, 1992) might have prevented the Salto river from capturing the Fucino Basin.

415 The drainage divide stability analysis (Forte and Whipple, 2018; see also Fig. S17; supporting 416 information) indicates an overall stability (as N°1-2-5-6-7 and 8, Fig. 8b) that is apparently 417 surprising as the Fucino is a large endorheic basin at 700 m of elevation. This stability could be 418 interpreted as due to the activity of the faults that keeps the divide high in elevation and contrast 419 the integration of the external rivers into the Fucino Basin. This is due to the fact that the faults 420 deformation rates are higher than the incision rates of the external rivers, pinning the location of 421 the drainage divide. The stability of the Fucino-Salto divide (divide 3; Fig.8b) is preserved by the 422 low stream power of the Salto R. since its upstream portion has not been reached by the erosion 423 related to the onset of the Quaternary regional uplift as evidenced by the very low values of 424 channel steepness (Fig. 8e) and of the local relief (Fig. 8c). The Fucino-Sangro stable divide (divide 425 9 and 10; Fig. 8b) may represent a snapshot of a stability referred to previous boundary 426 conditions. Indeed, the upstream Sangro R. valley is wide and hosts Lower Pleistocene fan deposits 427 scarcely incised (Miccadei et al., 2012), suggesting limited erosion.

In summary, morphometric observations of the Fucino Basin indicate that tectonic deformation rates are faster than the erosion rates. This can be also inferred by the fact that the inactive (since middle Pliocene) compressional structures are still shaping the landscape and by the overall stable divides. We suggest that the Salto-Fucino divide migration occurred in recent time (likely since late-Pleistocene, Centamore, 2004), but it has been halted by the Valle Majelama fan (Frezzotti and Giraudi, 1992). Since this upstream portion of the Salto River has not been reached by the erosion relative to the regional uplift, the divides tend to have a very low rate of migration.

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5.3. Schematic evolution of the Fucino Basin and changes in the direction of extension

437 Slip-directions, vertical offsets, Holocene slip rates and timing indicate an evolving pattern of 438 deformation around the Fucino Basin that we can summarize in two main stages shown in Fig. 9.

Stage 1 (Fig. 9a). The San Benedetto-Ovindoli, Tremonti and Velino-Magnola faults are the oldest structures, and likely formed during the late Pliocene (Cavinato et al., 2002), coeval with the onset of the Sulmona, L'Aquila and the Rieti basins (Cosentino et al., 2017). The kinematics of the Tremonti and Velino-Magnola faults were transtensive to dextral strike-slip (see also Cavinato et al., 2002), acting as transfer of the San Benedetto master fault. The coeval activity is well constrained by the oldest sedimentary sequence that is deposited at the junction between the San Benedetto and the Tremonti fault (Cavinato et al., 2002) (Fig. 9a). 446 Stage 2 (Fig. 9b, c and d). The San Benedetto fault laterally propagates (northward and 447 southward), reactivating older thrust faults, as the case of the Ovindoli-Celano structure (De 448 Vittorio and Faccenna, 1987). During this time, the direction of extension rotated from N240° to 449 N200° (Figure 9b). This new direction of extension (N200°) might have favored ruptures along new 450 fault planes oriented even WNW-ESE, such as the Piani di Pezza or Campo Felice, rather than 451 reactivating the deeper structures more oriented NNW-SSE, such as the Ovindoli or San Benedetto ones. Moreover, this new extensional field (N200°) favors reactivation with dip-slip kinematics of 452 453 the Velino-Magnola and Tremonti structures which strike instead from WNW-ESE to NNW-SSE 454 (Fig. 9b).

455 We suggest that the timing of the rotations in the slip-direction might have occurred around the 456 middle Pleistocene before the formation of western Piani di Pezza (PP1 site of measurement) and 457 Campo Felice faults that seem to present only the recent ~N200° slip-direction (see station CF3 and CF1 from this study, and Morewood and Roberts, 2000; Wilkinson et al., 2015; Fig S7 458 459 supporting Information). This is also in good agreement with the age of a sedimentary hiatus 460 between two sedimentary sequences within the Fucino Basin (Fig. 2), Pliocene-early Pleistocene 461 and since middle-late Pleistocene in age, related to two separated extensional events (Cavinato et 462 al., 2002).

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464 **5.4.** The Fucino Basin, the Central Apennines, and the role of the sediment loading

465 The faults that border the Fucino Basin behave differently with respect to the Central Apennines466 geological context. In particular:

467 1) The most recent N200° direction of extension only occurs in the northern sector of the Fucino 468 Basin. In fact, in the Central Apennines, slip-directions collected by previous studies, show a 469 main extension, oriented NE-SW (Faure Walker et al., 2010). Moreover, stress maps derived 470 from earthquakes and borehole breakouts indicate an overall direction for the minimum 471 horizontal sigma ranging between ENE-WSW to NE-SW (Montone et al., 2012, 1999). This also agrees with a second invariant strain rate from geodesy showing a principal direction oriented 472 473 NE-SW (Fig. 1a; D'Agostino et al., 2011). Similarly, in the southern Apennines, the current 474 direction of extension is oriented NE-SW (Montone et al., 2012, 1999). This NE-SW direction of 475 extension is also documented by previous slickenlines analysis (e.g. Papanikolaou and Roberts, 2007). For this region, the extensional direction over-imposed a previous one oriented NNW-476 477 SSE developed since late Miocene (Hippolyte et al., 1994).

The cumulative offsets across the faults located on the northern side of the Fucino Basin are
larger than other faults in the Central Apennines. The vertical throw along the Velino-Magnola
and Tremonti faults is ~3.5 km which is higher than the vertical offsets of San BenedettoPescina/Collarmele faults or L'Aquila, Sulmona and Rieti faults, that all range between 1.0 and
2.5 km (Roberts and Michetti, 2004).

483 These differences between the Fucino and the Central Apennines faults in terms of direction of 484 extension and cumulative vertical displacements could suggest that a local process is occurring at 485 the scale of the Fucino Basin. In fact, comparison between independent data such as vertical 486 geological displacement from this study, morphological displacement and the fault escarpments 487 height from literature (Benedetti et al., 2013; Schlagenhauf et al., 2011; Villani et al., 2015; Wilkinson et al., 2015), indicates a remarkable increase of the values from north to south, i.e. 488 489 towards the Fucino Basin (Fig. 10). The only exception to this pattern is the Velino vertical 490 geological displacement which is lower than the northern faults. In this context, we note that only 491 the Velino fault exhibits a gentle fault dip measured on the surface (<45°; see Table S1 and Fig. S3; 492 supporting information). The overall pattern shown in Fig. 10 strongly accords with Holocene slip-493 rates that also increases toward the basin, reaching maximum values along the Velino-Magnola 494 fault (Benedetti et al., 2013). Hence, our and previous data indicate that the northern structures of 495 the Fucino Basin are characterized by larger and faster displacement than the southern ones.

496 We propose that sedimentary (un)loading plays an important role in the internal dynamics of the 497 Fucino Basin, influencing the deformation pattern. On a first approximation, the sedimentary 498 (un)loading over the faults may be estimated assuming local Airy isostasy (Fig. S18; Supporting 499 information; England and Molnar, 1990). For the Fucino Basin, assuming a sediment thickness of ~1 km (Cavinato et al., 2002) with density of 1700 kg/m³, we estimate ~400 m of subsidence of the 500 501 San Benedetto hanging wall (Fig. 11 and Fig. S19; Supporting information). For the Magnola sector, 502 assuming ~1200 meters as a minimum estimation of thickness of the eroded material (Wise, 503 1985), we estimate maximum 1000 m of footwall uplift (Supporting information, Fig. S20). 504 Assuming flexural isostasy (see details in supporting information) and using elastic thicknesses (Te) 505 of 2, 4 and 6 km (cfr. with Te of 3.7 km; D'Agostino and McKenzie, 1999) we generally obtain, as 506 expected, lower values (see Figs. S21 and S22; Supporting information). For the hanging wall subsidence, we estimate ~369 (Te= 2 km), ~223 (Te= 4 km) and ~165 m (Te= 6 km) while for the 507 508 footwall uplift ~425 (Te= 2 km), ~150 (Te= 4 km) and ~81 m (Te= 6 km). These isostatic vertical

509 movements due to the (un)loading adjustment represent almost the 30% of the total geological 510 vertical slip.

The sedimentary loading may also produce a local rotation of slip-direction, from N240° to N200°. 511 512 According to Anderson regime (Anderson, 1951), the increase in vertical stress due to sediment 513 loading, which for the strike/oblique-slip is the intermediate stress (σ 2; Anderson, 1951), may be 514 enough to turn vertical stress into the principal one (Fig. 11b) (σ1; Anderson, 1951). This would 515 result in a re-organization of the stresses. As discussed in section 4.3, we suggest that the change 516 in the slip-direction, from N240° to N200°, and in kinematics has occurred during middle 517 Pleistocene, according to the hiatus related to an extensional phase and to the regional uplift 518 between two sequences within the Fucino Basin (Cavinato et al., 2002). This implies that the older 519 sequence 3 was related to an extensional event oriented N240°, while the younger sequence 4 520 was deposited during an extensional event oriented N200°. Based on the thickness of the two 521 sequences, we estimate that, since the onset of the deformation, the total vertical stress 522 incremented by ~15 MPa. According to this speculation, only ~4 MPa (estimated accounting only 523 the thickness of sequence 4; Cavinato et al., 2002) were needed to turn the kinematic for 524 strike/oblique- to dip-slip, as the case of the Tremonti fault (Fig. 11b).

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5.5. Feedback between tectonics, sediment loading and erosion

527 The feedback between tectonics and sediment (un)loading might have a fundamental impact on 528 the evolution of deformation. In extensional settings, hanging wall sedimentary load increases 529 faults slip (Hampel et al., 2009, 2007; Turpeinen et al., 2008) by up to 15% (Maniatis et al., 2009). 530 This base level lowering as a result of fault slip increase, might increase the river slope profile and 531 sediment production, resulting in an increased loading of the hanging wall. Conversely, in 532 compressional settings, the sediment load over the hanging wall may delay thrust slip (Ballato et 533 al., 2019; Hampel et al., 2009). Other studies reported that sediments or ice sheets (un)loading 534 influence or even trigger seismicity (Hampel et al., 2009). Erosional unloading may gently flexure 535 the upper crust, reducing the normal stress sufficiently to reactivate pre-existing structures, as 536 occurred in the New Madrid seismic zone (United States) (Calais et al., 2010; Hampel et al., 2007; 537 Hampel and Hetzel, 2006; Hetzel and Hampel, 2005; Steer et al., 2014; Vernant et al., 2013).

538 In this context, the development of internally drained basins, and therefore the resultant storage 539 of remarkable volumes of sediment, may exert a significant control on the kinematics of 540 deformation, even at the scale of the orogens (e.g. Royden, 1996; Willett and Beaumont, 1994). 541 Generally, endorheic-exorheic or exorheic-endorheic transitions are the result of a competition 542 between faulting, headward erosion and climate (e.g. Berry et al., 2019; D'Agostino et al., 2001b). Although this is theoretically applicable in several geological contexts, it is important to 543 544 differentiate the formation of significantly large endorheic basins at the scale of mountain wedges 545 and the formation of small intermontane endorheic basins, as for example the case of the 546 Apennines. At the scale of an orogen, where significant crustal shortening is occurring, the 547 development of topographic barrier in the frontal range may limit the precipitation behind, 548 significantly reducing fluvial incision. This, in turn, may develop large internally drained basins, 549 such as on Puna-Altiplano Plateau, the Tibetan Plateau, and the Tarim basin (e.g. Sobel et al., 550 2003). The accumulation of kilometers of sediments at high-elevation basin forces the 551 deformation to shift toward the margins of the orogeny, significantly enlarging the wedge. In turn, 552 this produces a lowering in slope, relief and therefore fluvial incision, favoring the preservation 553 through time of the endorheic configuration (Royden, 1996; Willett, 1999; Willett and Beaumont, 554 1994).

555 At smaller scales, such as the Apennines, where the shadow effect of precipitation can be 556 neglected, endorheic-exorheic or exorheic-endorheic transitions are a common process mainly 557 due to the competition between faulting and fluvial headward erosion. If fluvial incision is faster 558 than footwall uplift along a fault, the rivers might cut across the fault, capturing a basin drainage 559 system (Fig. 12a). If the slip or footwall uplift rate become faster than the fluvial incision, the basin 560 becomes endorheic (Fig. 12a). The surface processes are affected by climate variations, that locally 561 might be the main controlling factor in decreasing in the fluvial incision. In our study area, this 562 process may have occurred in the Campo Felice or Piani di Pezza basins, where wind gaps 563 (Forcamiccia and Valico Di Pezza; Figs. 12 and 4A) suggest a previous exorheic configuration (Fig. 564 12a; Giraudi and Frezzotti, 1997; Giraudi et al., 2011). In any case, the small internally drainage 565 condition, especially for intermontane extensional basins, is not permanent and it is commonly 566 interrupted by a new capture from a nearby catchment.

If at the onset of deformation, the fault slip rate is faster than the erosion rate, the basin will be endorheic (Fig. 12b). If the fluvial incision of adjacent basins becomes faster than the slip-rate, the basin might be captured (Fig. 12b). A second scenario to explain the endorheic-exorheic transition is the one associate to overspill processes (e.g. Geurts et al., 2019, 2018). This last model requires that the amount of water and sediment load accumulated within the internally drained basin rises to overflow the topographic barrier triggering the incision. In any case, in both scenarios (endorheic-exorheic transition due to headward erosion or overspill), this new externally drainage configuration might be permanent since, once captured, the increased discharge makes the fluvial incision rates much faster than vertical movement of the faults (D'Agostino et al., 2001b). In the Central Apennines an example of endorheic-exorheic transition is the Sulmona basin, where the Popoli Gorges cut through the Monte Morrone, linking the Sulmona basin with the Adriatic Sea (D'Agostino et al., 2001b; Geurts et al., 2019; Fig. 1 and 12b).

579 Interaction between erosion and faulting implies that during their evolution, the basins might be 580 characterized by both endorheic and exorheic phases, making the permanent endorheic nature, as 581 the case of the Fucino Basin, non-sustainable.

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5.6. Why is the endorheic nature of the Fucino Basin preserved?

We propose a conceptual model that considers the original structural setting of the Fucino Basin. We attribute the endorheic nature of the Fucino Basin to the coeval presence of master (San Benedetto) and transfer faults (Tremonti and Velino-Magnola). We propose that the presence of these two orthogonal faults favors the endorheic drainage stage since the onset of extension (Fig. 13).

In our model for the preservation of the internally drained configuration since the onset of extension and through time the two following conditions are strongly tied to each other: (1) the sediment load and (2) the combination of the two orthogonal sets of faults.

592 1) Our work indicates that sediment-related vertical geological displacements only 593 represent up to 30% of cumulative geologic throw across the San Benedetto-Pescina-594 Collarmele and the Magnola faults. This non-extensional vertical displacement, in turn, 595 increases the Fucino base level drop-down, enhancing erosion processes at higher 596 elevation surface. At the same time, the enlarged accommodation space developed by 597 faulting, allows to store more sediments overloading the faults hanging wall, inducing a 598 self-sustaining positive feedback process.

2) The coexistence of two orthogonal faults (Tremonti-Celano Aielli Vs San Benetto-Collarmele-Pescina) represents the main difference between the Fucino Basin and all the other extensional basins in the Central Apennines (Fig. 1). We suggest that this is the key for the preservation of the endorheic nature of the Fucino Basin through time. Our structural analysis shows a re-orientation along the transfer faults in the stress from N240° to N200°, from strike/oblique- to dip-slip. We proposed that this local change in the direction of extension and kinematic along the transfer faults might be related to the sediment loading. This increment of vertical stress may turn the vertical intermediate stress (σ 2) into the principal stress (σ 1; Anderson, 1951; Fig. 11). In turn, this kinematic changing along the original transfer faults favors the vertical displacement and subsidence on both master and transfer structures, self-triggering this process as described in point (1).

We suggest that the sediment-related additional vertical movement is one of the controlling factors for the permanent endorheic configuration of the Fucino Basin. It is worth to note that the Fucino Basin is located in a region that experiences both regional extension and uplift (San Jose et al., 2020). The uplift is fundamental to create, with the local footwall uplift, a topographic barrier against the nearby drainage basins.

In conclusion, we suggest that the competition between the sediment loading and the two orthogonal sets of faults that self-sustain each other with a positive feedback, could explain why the Fucino Basin remained endorheic through time. This increased subsidence prevents the basin to be captured by nearby catchments while the orthogonal set of structures that boarder the basin keeps the internal rivers to drain outside (Fig. 13).

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622 6. Conclusions

We presented a detailed structural analysis of the Fucino Basin, estimating the vertical geological displacement of the main faults surrounding the region and the expected effects on vertical displacement using both local and flexural isostasy. We propose a conceptual model to explain the peculiar and permanent endorheic nature of the Fucino Basin.

The vertical geological offset of the main faults ranges from 500 to 2500 m increasing from north to south toward the Fucino Basin. Our estimates indicate that up to 30% of these geological offsets are due to sediment (un)loading. We also suggest that the consequent significant increase in vertical stress may lead to a re-organization of the kinematics, from oblique- to dip-slip, especially along the transfer faults such as Tremonti and Magnola faults. This change in the kinematics produces a local rotation in the main slip-direction from N240° to N200°.

These findings allowed us to propose a conceptual model to better explain the endorheic nature of the Fucino Basin through time. We proposed that the combination of two orthogonal families of faults (San Benedetto and Tremonti) and the increasing slip, related to the sediment loading, prevents the Fucino Basin to be captured by nearby catchments. In turn, this process produces a

- 637 positive feedback increasing the sediment overloading the hanging wall, which in turn increases638 the slip of the fault, preserving the internally drainage through time.
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862 Figure Captions

863

Figure 1: a) Topographic map of the Central Apennines with schematic traces of normal and thrust faults b)
 Topographic map of the Central Apennines with the river network and the main rivers.

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Figure 2: Simplified geological map of the study area (see Fig. 1 for location) with in black the cross-section

868 869 traces. At the bottom, schematic geological sections modified after Cavinato et al. (2002) and from 'Foglio Geologico Sora 152' (Praturlon, 1968).

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Figure 3: Simplified structural map of the study area (see Fig. 1 for location) with the location of the
measurement sites. The black arrows indicate the direction of movement of the hanging wall with
respect to the footwall for each site. A merged stereoplot is shown for each fault with the hanging
wall slip-direction (blue arrows).

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Figure 4: A) Panoramic view of the Piani Di Pezza Fault. B and C) Detail of two slickenlines directions and
cross cutting relationship on PP2 measurement site (Fig. 3). D) Stereoplot of the fault planes and
slickenlines contour of PP2 measurement site.

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Figure 5: A) Panoramic view of the Magnola Fault. B) Detail of the M2 measurement site (Fig. 3). C) Detail of
 two slickenlines directions and cross cutting relationship on M2 measurement site (Fig. 3). D)
 Stereoplot of the fault planes and slickenlines contour of M2 measurement site.

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- Figure 6: A) Panoramic view of the Tremonti Fault. B) Detail of the TRM4 measurement site (Fig.3). C) Detail
 of two slickenlines directions and cross cutting relationship on TRM4 measurement site (Fig. 3). D)
 Stereoplot of the fault planes and slickenlines contour of TRM4 measurement site.
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Figure 7: A) Panoramic view of the Gioia dei Marsi Fault. B) Detail of the GdM1 measurement site (Fig. 3). C)
 Detail of two slickenlines directions and cross cutting relationship on GdM3 measurement site (Fig. 3).
 D) Stereoplot of the fault planes and slickenlines contour of GdM3 measurement site.

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Figure 8: a) Topographic map of the study area (left) and two swath profiles (right). The white boxes
indicate the swath profile locations and the numbers the divide sections included in the divide
stability analysis. b) Plot showing the results of the drainage divide stability analysis following Forte
and Whipple (2018). c) Local relief map with 5 km of average circular window. d) Slope map. e) Map of
the along channel variation of *k_{sn}*.

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Figure 9: Schematic evolutionary model from late Pliocene to Holocene, inspired by Cavinato et al. (2002).

Figure 10: Simplified structural map of the study area (see Fig.1 for location) with the fault labeled based on
a) escarpment height (short-term displacement; see Puliti et al. (2020) for detail) from previous

902 studies (see main text for references), b) Morphological displacement (mid-term displacement; see

- Puliti et al. (2020) for detail) from previous studies (see main text for references) and c) geological
 throw (long-term displacement; see Puliti et al. (2020) for detail) from this work. In the lower plot
 displacements over faults.
- 906

Figure 11: The role of the sediments in the deformation. a) Sediments thickness from Cavinato et al. (2002)
 on the left and subsidence calculated using local isostasy model from Molnar and England, (1990). In
 the lower part topographic section: current topography, bedrock depth and expected bedrock depth
 without sediment loading. b) Kinematic changing and stress re-organization on the transfer fault
 (Tremonti Fault).

- 912
- 913 **Figure 12**: Conceptual model for endorheic-exorheic and exorheic-endorheic transition for small
- 914 intermontane basins. a) Evolution from exorheic to endorheic with a panoramic view of Campo Felice
- basin. b) Evolution from endorheic to exorheic with a panoramic view of Sulmona basin.
- 916
- 917 Figure 13: Schematic and conceptual model for the endorheic preservation through time of the Fucino918 basin.
- 919

Figure1.



Figure2.



Figure3.





Figure4.





Figure5.

Figure6.

Figure7.

Figure8.

Figure9.

Figure10.

Figure11.

a) Increasing in vertical displacement

Strike/Oblique-slip

Normal dip-slip

Figure12.

Figure13.

Schematic Structural evolution of the Fucino Basin

-Increasing in sedimentation -Kinematic changing from oblique- to dip-slip on the transfer fault

