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The Vaiont landslide: re-assessment of the evidence leads to rejection of the consensus

Abstract There appears to be a clear general consensus in the literature regarding four critical issues that define the problem of the October 1963 Vaiont landslide and its behaviour that are central to the disaster: (1) the 1963 failure was a reactivation of an ancient landslide; (2) failure took place along thin clay seams (already at residual strength); (3) the sliding surface had a ‘chair’ shape with a (sub)horizontal base; and (4) failure was triggered by inundation of the toe of the slide mass by rising reservoir levels. The key to understanding the Vaiont landslide is the failure surface geometry, which was controlled by the structural geology. It now appears that the so-called chair structure (that was assumed to define the shape of the failure surface) does not exist, and without it, the first consensual point is untenable, and the fourth may not contain the whole truth. We have systematically re-examined the published evidence and undertaken our own new research in order to test the logical and geotechnical validity of the four elements of the consensus. Glacial processes can account for the pre-failure morphology of the landslide site; the clay seams must therefore have been at peak shear strength as there was no ancient landslide. Tectonic processes can account for the failure surface geometry, which does not have a ‘chair’ shape, as well as small-scale structures; and rainfall appears to have been an essential element in the initiation and development of the landslide. Our findings largely contradict the consensus position and thus form the basis of a new overarching hypothesis for the landslide that should account for all of the observed and known features, events and data.

Keywords Vaiont · Landslide · Rock slide · Geology · Geotechnics · Failure surface geometry

Introduction

The story of the Vaiont dam, reservoir, landslide, disaster and subsequent scientific investigations into the landslide is well known in the scientific and engineering literature and has even featured in several television programmes including a joint Italian-French ‘docudrama’ film ‘*Vajont: La diga del disonore/La Folie des hommes*’ released in 2001. In contrast to the populist note of this film, the 50th Anniversary of the Vaiont (or Vajont) landslide (Fig. 1a) was marked by a conference held in Padua in October 2013. The conference aim was ‘*to prompt a debate among all the scholars of the world on the event of 9 October 1963 and on its repercussions on the evolution of technical-scientific knowledge and of international legislation*’ (Genevois and Prestininzi 2013, p.vii).

Notwithstanding the many contradictions and debates that have persisted through the decades of research into the landslide, located in northeast Italy (Fig. 1b), the scientific and geotechnical arguments about the nature of the materials involved and the mechanics of the landslide have broadly converged towards a general consensus, the elements of which are identified in Table 1. Through the contributions to the 50th Anniversary conference, this consensus appeared to be widely held and unchallenged.

Despite this consensus, no explanation of the 1963 landslide has ever been able to account for both its occurrence *and* the observed large velocity and displacement.

The hypothesis of this paper is that the consensus position is not tenable. In this paper, we examine these elements in order to identify the critical evidence and arguments that led to the consensus, as well as the contradictory issues that have thus far prevented any holistic explanation of the landslide. In essence, the most obvious problem that presents itself is an uncertainty about the mechanisms that caused the supposed pre-existing landslide to move so quickly from a position of rest, as the sliding surface of a pre-existing landslide along clay seams should have exhibited the shear strength properties of a sheared clay. Experience elsewhere suggests that this strength cannot be reduced further without invoking one of a range of postulated, but never widely experienced, phenomena. We have approached the problem from the perspectives of geomorphological setting, geological probability and geotechnical argument to determine which are the critical assumptions in the consensus that lead to this impasse.

The starting point for our study was to question whether the Vaiont landslide actually had a ‘chair’ structure. Others have recently independently established that it did not (Bistacchi et al. 2013) and indeed could not (Petronio et al. 2016) have had a ‘chair’, so we will examine this issue and the consequential implications that arise from it, i.e. (i) that there was no ancient landslide; (ii) that submergence of the toe by the impounded reservoir was not the sole cause of the failure; and (iii) that a new hypothesis can be proposed which should allow the entire event to be explained in terms of the available evidence and basic geotechnical theory. The aim of this paper is therefore to review and, if necessary, modify or replace the elements of the consensus to provide a new framework for attempting to explain the occurrence and behaviour of the Vaiont landslide.

Any overarching explanation of the Vaiont landslide must account for many disparate and seemingly contradictory observations, measurements and other evidence. We have gone back to original data and other contemporary evidence (e.g. the extensive collection of photographs taken by Edoardo Semenza and colleagues: Masè et al. 2004) as well as some more recent re-analyses of original data using modern technologies. In doing so, we identified several features of the development and occurrence of the landslide for which published accounts demanded further consideration. Some of these features or events had previously been dismissed as not important for any explanation; others may have interpretations different from those previously published and perhaps largely accepted. Indeed, most of the critical information had been either identified or correctly hypothesised (at least broadly) shortly after the landslide, but became obscured by what were then convincing alternative interpretations. These features and events are listed in Table 2 in approximate chronological order of consideration by contemporary studies. The overall structure of this paper therefore also follows an approximate historical timeline of evidence acquisition and observed events.



Fig. 1 a The Vaiont landslide as seen from Casso in 2011. The top of the dam can be seen at the lower right corner. Photo: ENB. b Location of Vaiont in the Dolomites north of Venice (Venezia), Italy. The grey/white shading identifies the administrative regions

Features and events

Geomorphological context—F1

A fundamental question is how and why (and therefore when) the Vaiont gorge formed, since this may provide some clues regarding

long-term changes to the geotechnical competence of the valley slopes. Related to this is the question of how the north slope of Mt. Toc developed a morphology that, prior to 1963, was consistent with a landslide including, for example, the raised masses of the Pian della Pozza and the Pian del Toc (Fig. 2) within a broad bowl-

Table 1 Elements of the existing consensus relating to the Vaiont landslide, showing reference codes used in this paper

| Element of consensus | | References |
|----------------------|---|---|
| C1 | The failure was a reactivation of an ancient landslide. | Guidici and Semenza (1960) ¹ , Semenza (1965), Voight and Faust (1982), Hendron and Patton (1985), Belloni and Stefani (1992), Nonveiller (1992), Tika and Hutchinson (1999), Mantovani and Vita-Finzi (2003), Ward and Day (2011), Superchi (2012), Hungr and Aaron (2013), Paronuzzi and Bolla (2012, 2015), Paronuzzi et al. (2016), Crosta et al. (2016), Zhao et al. (2016), Pasuto (2017) |
| C2 | Both the prehistoric landslide and the 1963 landslide took place along thin clay seams in a limestone mass and that these were at, or near, residual strength prior to 1960. | Semenza (1965), Skempton (1966) ² , Müller (1968) ³ , Voight and Faust (1982), Hendron and Patton (1985), Tika and Hutchinson (1999), Ward and Day (2011), Hungr and Aaron (2013), Zhao et al. (2016) |
| C3 | The shape of the sliding surface followed a folded rock structure that comprised a subhorizontal lower part and a steeply inclined upper part, commonly referred to in the literature as a 'chair' shape. | Guidici and Semenza (1960), Müller (1964, 1968), Semenza (1965), Mend (1966) ⁴ , Skempton (1966), Broili (1967) ⁵ , Romero and Molina (1974), Chowdhury (1978), Hendron and Patton (1985) ⁵ , Belloni and Stefani (1992), Tika and Hutchinson (1999), Kilburn and Petley (2003), Hungr and Aaron (2013) ⁶ , Paronuzzi and Bolla (2012, 2015), Boon et al. (2014) ⁷ , Paronuzzi et al. (2016) ⁵ , Crosta et al. (2016) |
| C4 | The trigger for the failure was a major loss of stability due to inundation of the toe of the slide mass by the impounded reservoir. | Kiersch (1964), Semenza (1965), Chowdhury (1978), Hendron and Patton (1985) ⁸ , Ward and Day (2011), Delle Rose (2012), Paronuzzi and Bolla (2012, 2015), Paronuzzi et al. (2016) |
| C1–C4 | All of the above. | Hutchinson and Kwan (unpubl.1986), Hutchinson (1987) ⁹ , Alonso et al. (2010), Alonso and Pinyol (2010), Pinyol and Alonso (2010), Del Ventisette et al. (2015) ¹⁰ , Havaej et al. (2015) ^{11,12} , Wolter et al. (2016) |

¹ They suggested the presence of an ancient landslide prior to the failure event

² Due to flexural slip rather than a previous landslide

³ Not as 'residual strength' but as 'unusually low friction values' (ϕ_p)

⁴ Less extreme than some versions

⁵ The 3D 'bowl' shape was acknowledged

⁶ Similar to Broili (1967)

⁷ But recognised the limitations of this assumption

⁸ Rainfall influences are analysed and acknowledged

⁹ C3 implicit—referred to as 'strongly non-circular slide'

¹⁰ C4 'seemed to be'—not definitively their position

¹¹ But they refer to Bistacchi et al. (2013)

¹² Particularly that reservoir level affected the rate of displacement

Table 2 Features and events that constitute critical evidence for understanding the landslide

| | Feature or event (with corresponding consensus element) | Reference(s) |
|----|---|---|
| F1 | The pre-failure geomorphology of Mt. Toc. (C1) | Kiersch (1964) Müller (1964) Semenza (1965) Hendron and Patton (1986) Wolter et al. (2014, 2016) Pasuto (2017) |
| F2 | Geological outcrops recorded by Edoardo Semenza in 1959 and 1960: (i) cataclastic rock at the 'East Wall' of the Vaiont gorge; (ii) the Colle Isolato and the river sediments beneath this unit; (iii) cataclastic rock confining the western branch of the Massalezza Ditch above the Pian della Pozza. (C1) | Guidici and Semenza (1960) Semenza (1965, 2001, 2010) Broili (1967) |
| F3 | Underlying geological structures and failure surface characteristics (includes borehole evidence). (C2, C3) | Kiersch (1964) Müller (1964, 1968) Semenza (1965, 2001, 2010) Caloi (1966) Broili (1967) Hendron and Patton (1985) Belloni and Stefani (1992) Bistacchi et al. (2013) Ghirotti et al. (2013) Massironi et al. (2013) Boon et al. (2014) Petronio et al. (2016) |
| F4 | The 1960 landslide. (C4) | Semenza (2001, 2010) |
| F5 | The effect of reservoir levels and rainfall on pore pressures in the slopes (includes piezometer data). (C4) | Müller (1964) Hendron and Patton (1985) |
| F6 | The progressive eastward development of the failure surface (includes seismic evidence). (C1, C2, C3, C4) | Müller (1964) Caloi (1966) Belloni and Stefani (1992) Delle Rose (2012) |
| F7 | The acceleration, velocity and displacement of the main failure event (includes seismic evidence). (C2, C3) | Müller (1964, 1968) Delle Rose (2012) Zhao et al. (2016) |
| F8 | The retrogressive development of both the eastern* and western scars following the main failure event (photographic evidence). (C2, C3) | *Semenza (1965, 2001, 2010) *Hendron and Patton (1985) Wolter et al. (2016) |

shaped mountain slope (C1). In 1928 and again in 1958, examinations of the slopes of the Vaiont valley along the length of the planned reservoir by Prof. Giorgio Dal Piaz found that, with only a few localised exceptions, there were no concerns regarding the stability of either side and particularly of the left (southern) valley slopes (Bozzi et al. 1964, cited in Hendron and Patton 1986). The internal geology of the south side of the valley and Mt. Toc was suggested to be sufficiently complex to have resulted from a large mass movement (Rossi and Semenza 1965) but Broili (1967) reported that the intense folding of the rocks was tectonic and pre-existed any possible ancient landslide, recently confirmed by Massironi et al. (2013).

It has previously been established that (i) the Vaiont valley was glaciated (Castiglioni 1940, cited in Wolter et al. 2016) and (ii) during the Last Glacial Maximum (LGM), the Piave Valley glacier filled the Alpi and Belluno basins south of Longarone with an upper surface elevation of around 1100 m (Castiglioni 2004). Further, Semenza (2001, 2010) reported studies undertaken in 1951 and 1961 identifying that the Sant'Oswaldo Pass between Erto and Cimolais (Fig. 3) comprised a landslide mass, the summit of the pass being at 828 m elevation with the valley further west infilled with up to 50 m of permeable alluvial sediments over an impermeable base at 750 m, and that before this

landslide the Cimoliana stream 'must have' flowed west from Cimolais past Erto and down the Vaiont valley to Longarone, instead of south from Cimolais as it does now. According to this interpretation, the Vaiont gorge must be of relatively great age. Indeed, all previous accounts have assumed that the Vaiont river has always flowed towards the west (e.g. Wolter et al. 2016). Kiersch (1964), however, suggested that the 'inner gorge' may be less than 18,000 years old, its rapid incision producing rock walls not fully adjusted to the stress relief. Others have also suggested a late glacial or immediately post-glacial origin for the gorge (Alonso et al. 2010; Wolter et al. 2014).

We agree with Kiersch's assessment of maladjusted rock stresses arising from a rapidly incised gorge—but resulting from a post-glacial reversal of drainage along the Vaiont valley. If the gorge is ignored, the morphology of this valley before 1963 is directly comparable with that of the present Zemola valley immediately north of Erto (Fig. 3), i.e. reflecting a glacially enlarged headwater catchment separated from the Piave valley by a ridge somewhat higher than its present 880 m elevation.

Even following glacial deepening of the Erto Basin, the Vaiont river upstream of Erto is at the same elevation as Cimolais. Indeed, the Cimoliana valley appears to open out in the downstream direction from Cimolais and the scale of this valley suggests that its floor elevation would always have been

Mt. Toc (Fig. 3), as indicated by Pasuto (2017) who described recently exposed rock striations above Casso village that demonstrate such ice movement. The morphology of the northern slope of Mt. Toc would therefore result from the geometry and mechanics of this glacier movement, with erosion of the upper slopes being strongly determined by the bedding, although some mass movements undoubtedly occurred. Indeed, several earlier publications describe moraine deposits on slopes above the inner gorge (Kiersch 1964), 500 m east of the Pian della Pozza (Müller 1964), and Semenza (1965), p.G28 in Hendron and Patton (1985) noted that ‘The presence of morainic deposits remains problematic’. Hendron and Patton (1986) suggested that some of the surface depressions on the north slope of Mt. Toc, particularly the large elongated feature at the upslope side of the Pian della Pozza, may have been glacially scoured. If the surface of the ice was at around 1100 m elevation (or possibly somewhat higher this far north) then it seems likely that the thick sediments found by Semenza (2001, 2010) in 1960 along the western branch of the Massalezza stream above the west side of the landslide at around 1000 m elevation probably comprised a lateral moraine. The ‘partly cemented breccia’ described by Hendron and Patton (1986) from an infilled pre-1960 tension crack, which they suggest could be less than 2000 years old, may therefore derive from this moraine sediment.

The area just east of the Vaiont slide site known as ‘La Pineda’ includes a ‘well-preserved layer of glacial deposit’ more than 20 m thick beneath 90 m of deposit from the post-glacial Pineda landslide (Fig. 3). This glacial deposit comprises delta sediments thought to be associated with a proglacial lake, covered by moraine and then colluvium (Pasuto 2017). Blockage of the Sant’Osvaldo valley by till and/or the large landslide would have impounded meltwater within the Erto Basin, creating a significant lake. Localised erosion by the overflowing Piave glacier of a notch through the ridge opposite Longarone would then have been exploited by meltwater to form the Vaiont gorge as a new drainage path from the basin. According to these interpretations, the gorge formed very rapidly during the late Pleistocene/early Holocene due to very high volumes of meltwater.

The implication is that a significant thickness of material may have been removed from Mt. Toc by glacial action prior to formation of the gorge, thus reducing the likelihood of very large-scale landsliding. The ‘old headscarp’ noted by Hendron and Patton (1986) above the 1963 western headscarp may result from a very shallow landslide or from erosion by the glacier. However, stress relief joints probably enhanced rainwater penetration into lower stratigraphic beds with concomitant karstic development and more rapid groundwater circulation. Thus, instability was promoted once the gorge was cut.

Semenza’s geological studies—F2

Most of the evidence for the Vaiont landslide being a reactivation of a pre-existing failure (C1) came from Edoardo Semenza, but was subsequently supported by independent assessments of the geomorphology and geology of the site. Semenza began his investigations of landslides in the Vaiont valley in August 1959. He identified several features (F2) that led him to conclude that much of the left bank of the gorge comprised an ancient landslide of perhaps several tens of millions of m³ in volume (Semenza 2001, 2010, Chapter 6). His critical evidence comprised: (1) geological characteristics where the Col Tramontin Fault met the ‘eastern wall’ on the left bank of the gorge upstream of the Massalezza Ditch; (2) the identification of the Colle Isolato on the right bank

of the gorge opposite the Pian della Pozza; and (3) the morphology and geology of the upper Mt. Toc slopes, particularly the western branch of the Massalezza Ditch. We examine these in turn.

1. At the so-called eastern wall of the gorge (upstream end, left bank; Fig. 2), Semenza found a layer of what he described as ‘extremely fractured rocks (cataclasts)’. He wrote: ‘It was logical to think that these cataclasts would also run under all the northern wall’ (2010, p. 61) and interpreted them as being the failure surface of an ancient landslide. However, these mylonites/cataclasts were not found any further towards the dam (Broili 1967), and borehole S3 that was drilled in 1960 (Fig. 2) did not reveal any such rocks of any thickness at the depth of the 1963 failure surface (also noted by Semenza 1965; see below). Hendron and Patton (1985) suggested the possibility of ‘tectonic faulting’ in the form of a thrust along the bedding creating a weakness at the position indicated by Guidici and Semenza (1960) could give the impression that the 1963 landslide was a reactivation of an old slide—although Hutchinson and Kwan (unpublished manuscript 1986) noted that if it was a tectonic shear then it may have recovered strength by 1960. Nonveiller (1992) also assumed that the 1963 landslide was a reactivation of a prehistoric slide but nevertheless with progressive failure taking place. The location of the ‘eastern wall’ outcrop observed by Semenza met the line of the Col Tramontin Fault. It seems reasonable to suggest that movements along this fault zone could have disrupted adjacent bedding to differing degrees resulting in variable thicknesses and distributions of fault gouge, especially if part of the bedding plane did coincide with a local thrust fault. Therefore, the observed ‘cataclasts’ do not constitute evidence of an ancient landslide.
2. The Colle Isolato comprised a mass of rock on the right side of the gorge (Fig. 2), resting on horizontal layers of river gravels that infill an older, narrower ‘inner’ gorge (Semenza 2001, 2010). It was structurally discordant with the in situ rock of the right side but consistent with the left side. Semenza’s interpretation of all the visible details was that this mass was probably a remnant of the toe of the ancient landslide that slid from the left side, crossed (and blocked?) the pre-1963 gorge and came to rest when it hit the right side of the gorge. However, the Colle Isolato deposit appears to have buried river gravels as observed and photographed by Semenza, i.e. it must post-date an earlier blockage and infilling of an older inner gorge. Neither the age nor indeed the extent of the gorge realignment affects the key point, which is that the landslide that formed the Colle Isolato does not necessarily constitute evidence of a large ancient landslide coincident (at least in part) with the extent of the 1963 failure. Broili (1967) noted that borehole S1, drilled in 1960 (Fig. 2) revealed a layer of silt and sand below the Pian della Pozza, perhaps 100 m higher than the 1963 failure surface, that could have resulted from a large slide of perhaps 30–40 million m³. This, or a smaller slide from the face of the otherwise intact left wall of the gorge (e.g. Fig. 17 in Wolter et al. 2016; Broili suggested ~ 9 million m³), could have formed the Colle Isolato. We suggest the latter to be the more likely case (Fig. 4)
3. Immediately above the upper outline of the 1963 landslide, the Massalezza Ditch branched with tributary valleys (sometimes referred to as ‘gullies’) extending upstream diagonally up the upper northern face of Mt. Toc towards the southwest and southeast (Fig. 2). In July 1960, Semenza examined the western branch and found the upslope (with respect to Mt. Toc) side of the stream valley to comprise exposed beds of the Fonzaso Formation, dipping

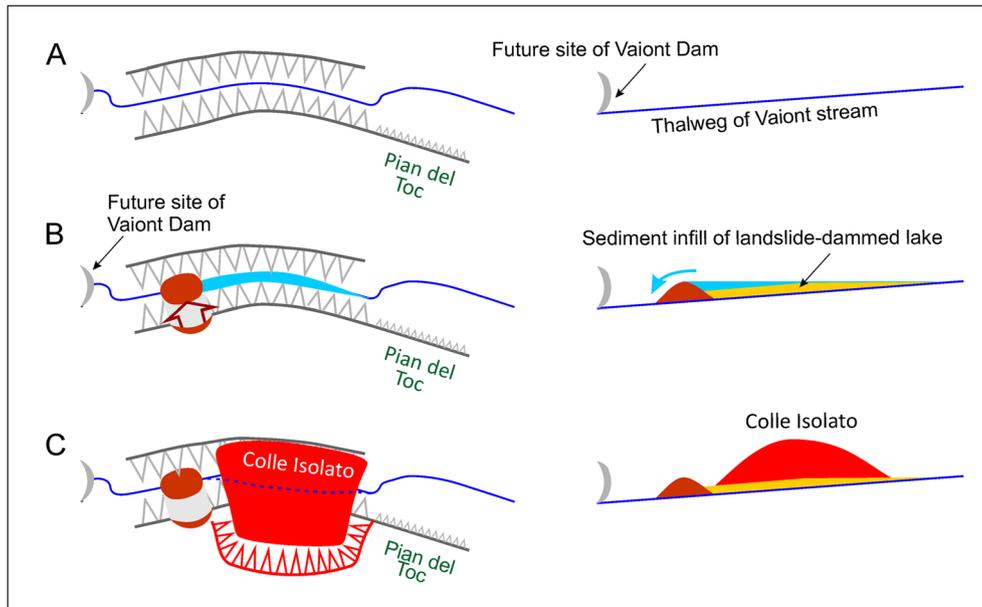


Fig. 4 Hypothetical explanation for the Colle Isolato and its apparently buried river gravels, showing plan and river long profile views. The diagrams are not to scale, and the post-1960 location of the dam is shown for context only. The grey triangles indicate very steep (probably near vertical) cliffs of the inner gorge. **a** Early alignment of the inner gorge upstream of the dam. **b** A small landslide from either side of the gorge (south side shown) blocks the inner gorge, which then infills upstream of the blockage with fluvio-glacial sediment as the river flows over (then through) the landslide dam. **c** A much larger landslide from the south side completely buries the old inner gorge and causes the river to cut a new course through the slide mass slightly further south (dotted blue line). The source area of the landslide shapes the outer gorge downslope of the Pian della Pozza, including the site of the November 1960 landslide, and the slide deposit forms the Colle Isolato

northwards at 40°, with the downslope side being defined by a ‘remarkable thickness’ of mylonites and cataclasts that he identified as part of the ancient palaeoslides (Semenza 2001, 2010). He also noted thin, highly fractured layers of ‘Fonzaso-Soccher’ rocks on the top of these sediments. We suggested previously (F1 in Section 2.1) that the observed pre-failure morphology of the north side of Mt. Toc can be explained in terms of glacial and post-glacial processes and events. That explanation provides an alternative interpretation for these sediments at Massalezza Ditch as being lateral moraine generated by the Piave glacier eroding the top of the ridge into the Vaiont valley. The ‘Fonzaso-Soccher’ rocks may be locally displaced by the ice movement or, perhaps more likely, the result of post-glacial shallow landsliding from the oversteepened slope above.

On the basis of our reviews of F1 and F2, we arrived at an interim conclusion that there was no ancient landslide and that the 1963 event was a first-time failure of the northern slope of Mt. Toc (c.f. C1). We will show later that this working hypothesis is consistent with all other elements of our investigation.

Geology and the failure surface—F3

Stratigraphic succession

The stratigraphic context of the Vaiont landslide is a sequence of Jurassic–Cretaceous limestones of varying lithologies. Table 3 presents a summary of the stratigraphy as presented in more recent published accounts. Most of the failure surface coincides with bedding planes within the Fonzaso Formation, mostly near the base of the upper unit of the Fonzaso Fm. according to recent seismic studies (Petronio et al. 2016). However, the presence of

thin clay layers between the limestone beds was initially a source of some debate. Broili (1967) and Müller (1968) both concluded that there was no clay but, instead, occasional very thin smears of pelitic materials with < 20% content of clay minerals, and that these intercalations played no significant role in the landslide. Hendron and Patton (1985) later confirmed both the presence and stratigraphic continuity of clay interbeds within the Fonzaso and Calcare di Socchère Formations, particularly from the basal failure surface upwards (C2). These layers (e.g. Fig. 5) sometimes exceeded 150 mm in thickness and comprised up to 50–80% clay minerals, dominantly calcium montmorillonite. This mineralogy suggests that the clay interbeds probably result from periodic, though distant, volcanic eruptions during deposition of the limestone sequence (Bernoulli and Peters 1970; Bromhead 2013).

We are satisfied that the issue of the clay layers was fully resolved by Hendron and Patton (1985), not least because we have seen such layers in situ in the failure surface of the landslide and, more clearly, in the road tunnel adjacent to the dam (Fig. 5). Furthermore, given that there are interbedded clays within the limestone formation that failed, it would be expected that failure of the slope would develop primarily along these layers because they are so much weaker even than friction along cracks through the limestone. This effect of shear stresses exploiting the weakest part of a soil or rock profile is well known and expected according to geotechnical theory.

Table 1 cites a wide range of published studies that consider the clay layers involved in the failure to have been at residual strength (C2) associated with the very large ancient landslide (C1), or at ‘reduced strength’ for other reasons such as flexural slip along bedding associated with tectonic deformations (e.g. Skempton 1966). Boon et al. (2014) presented a comprehensive

Table 3 Indicative stratigraphy of the south side of the Vaiont valley (after Bistacchi et al. 2013 and Ghirotti et al. 2013)

| Age | Stratigraphic name | Thickness (m) | Description | Context |
|--------------------------------------|---|---------------|--|--|
| Upper Cretaceous-Lower Palaeocene | Scaglia Rossa | ~ 300 | Red coloured layered marly limestones and marls | Not present on the left side of the Vaiont valley |
| Cretaceous | Calcare di Soccher, or Soccher Limestone ² | 150+ | Layered ¹ marly and cherty limestones Massive grey, red or greenish marly limestones | Landslide mass |
| Upper Jurassic (Tithonian-Oxfordian) | Rosso Ammonitico | 0–15 | Fossiliferous nodular micritic limestone | |
| Upper Jurassic (Oxfordian-Callovian) | Fonzaso Formation | 10–40 | Layered cherty micritic limestones with intercalated green clay layers ~ 5–180 mm thick Layered cherty micritic limestone | Landslide mass—the shear surface formed along the clay layers in this unit ³ In situ bedrock below the landslide failure surface |
| Middle Jurassic (Dogger) | Vajont Limestone | 350–450 | Massive resedimented oolitic limestones | |

¹ Implies thinly bedded

² Also known as the Biancone Formation (Ghirotti et al. 2013)

³ Petronio et al. (2016) interpreted new seismic data as showing the failure to have occurred at, or very close to, the boundary between the upper and lower units of the Fonzaso Formation

list of shear strength values for the clay layers that have been measured, derived from stability analyses or simply estimated for modelling purposes. Many of these fall in the range $\phi_r' = 6\text{--}12^\circ$, where ϕ_r' indicates the residual angle of shearing resistance of typical clays, although several others are only slightly less than 20° . This is higher than residual strength but still significantly lower than relevant peak strength which is probably around $\phi_p' = 22\text{--}26^\circ$ as found in clays in similar contexts to Vaiont (Petley 1999, cited in Kilburn and Petley 2003). The consensus position C2, and to some extent also C1, arises from the assumptions regarding the failure surface geometry (C3). Our examination of C3 leads to an argument that negates both C2 and C1.

Tectonic setting

Massironi et al. (2013) presented a new synthesis of the origins of the geological characteristics of the Vaiont valley and

landslide site based on field investigations, DTMs generated from photogrammetric analysis and LiDAR imagery. Figure 6 summarises the major tectonic structures that define the structural context of the site, which has been subjected to three significant stages of tectonic development. In summary: Stage 1 involved NE-SW to NNE-SSW compression, forming the eastward-plunging regional syncline to the north of the Vaiont valley identified as the Erto Syncline and producing the major thrusts immediately to the north (Mt Borgà and Spesse thrusts: Massironi et al. 2013). This stage tilted the Vaiont beds to form what is now the northern slope of Mt. Toc. Stage 2 comprised NW-SE compression that formed the regional anticline around 5 km southeast of the Vaiont valley and folded the Vaiont beds at Mt. Toc into a steeply plunging syncline (i.e. the Massalezza Syncline). This stage probably also produced the major N-S striking steep reverse faults (the Col delle Tosatte and Croda Bianca faults in Fig. 6) that converge downwards either side of

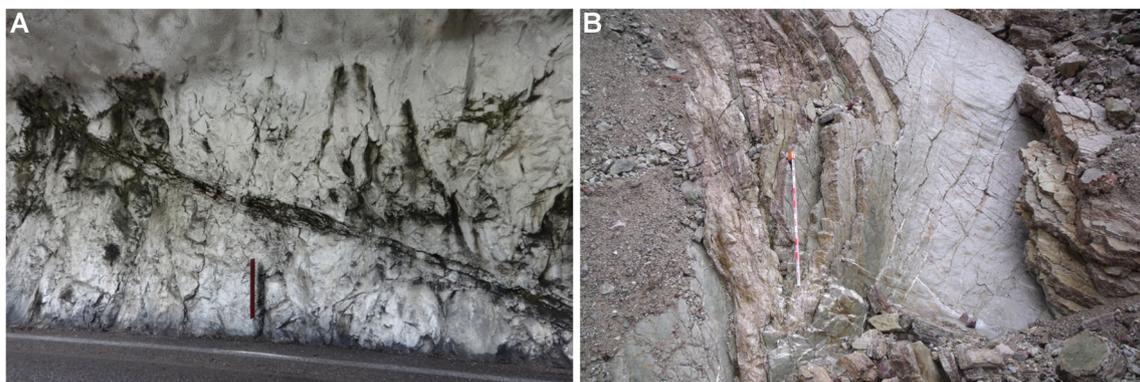


Fig. 5 a Interbedded clay layer within the Biancone Formation in the unlined wall of the road tunnel adjacent to the Vaiont dam (eastward dip to the right, the roof arch is shotcreted). Photo: ENB. b Thin beds of limestone and clay on the failure surface of Vaiont landslide just west of the Massalezza Ditch (see Ghirotti et al. 2013 for details of this location). The measurement scale is 1 m long. Photo: APD

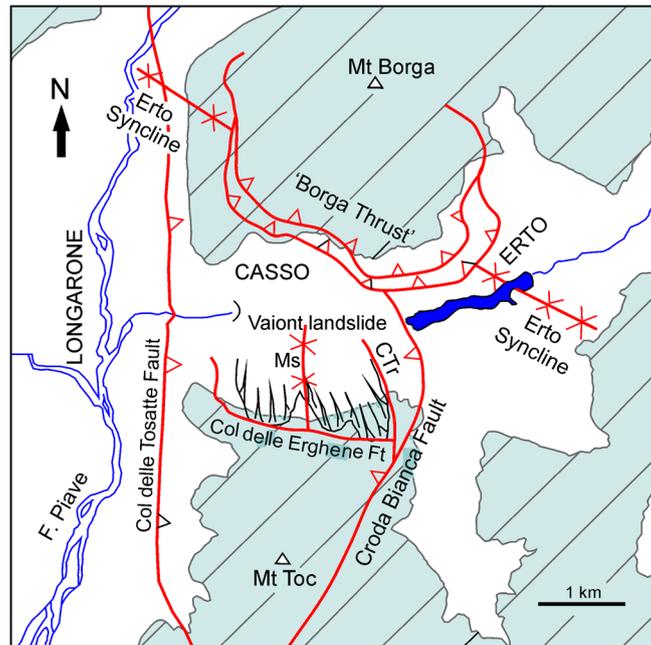


Fig. 6 Structural geology of the setting of the Vaiont landslide (8 km × 8 km area). Directions of fault plane dips are indicated. 'CTr' = Col Tramontin Fault, 'Ms' = Masalezza Syncline. The upper scar of the Vaiont landslide is marked by slope hatching. Land over 1100 m elevation is shaded. Based on Riva et al. (1990)

what became the 1963 landslide (Massironi et al. 2013). Stage 3 involves N-S compression which, although not impacting on the structures of the Vaiont valley, is ongoing in this region (Dr F. Podda, Università degli studi di Trieste, pers. comm., 2012).

Both sides of the 1963 failure surface display a variety of distinct patterns of structural features that have not changed despite some modifications by minor rockslides and rockfalls since 1963 (Wolter et al. 2014). The failure surfaces are sometimes fully planar, sometimes generally planar but with meso-scale undulations (tens of metres in lateral extent with amplitudes up to a few metres), and sometimes very complex where the in situ beds are locally highly deformed at the scale of ~ 1 m. Massironi et al.'s (2013) account presents detailed geometric characteristics of structures associated with interference between the two compression fields as well as flexural slip processes. This means that Broili's (1967) assessment that the structural irregularity of the failure surface was of tectonic origin and not due to an ancient landslide appears to have been substantially correct.

The critical conclusions from Massironi et al.'s study, consistent with our own, are that the tectonic movements and deformations, across as well as along bedding, probably controlled and even promoted formation of the failure surface along different clay-rich beds within the stratigraphy, and that the highly variable patterns of tectonic fracturing throughout the in situ rocks also predisposed the northern slope of Mt. Toc to instability. The development of the 1963 landslide and the form of its failure surface can therefore be fully attributed to the inherent geology (lithology and stratigraphy) and tectonic history of the site. The large-scale form of the failure surface, arising from the stage 1 compression outlined above, now requires detailed consideration (C₃).

Shape of the concealed failure surface: the 'chair'

Many of the early assumptions—and thus interpretations—were based on field geological observations and mapping, which were reinforced by subsequent information such as borehole cores that had poor rates of recovery (< 30% even from intact rock below the failure surface: Hendron and Patton 1985), from a stratigraphic sequence about which there was far less understanding at that time. The failure surface of the 1963 landslide was described as having a 'chair' shape (C₃) on the basis of observations of exposed rock faces in the Piave valley at Longarone. Indeed, this structure is clearly visible in the left wall of the valley immediately south of the downstream end of the Vaiont gorge (Fig. 7) and can also be seen at a higher elevation in the right wall. It had been assumed that this structure, relating to the Erto Syncline, persisted upstream along the gorge though with the subhorizontal 'seat' dipping eastwards at up to 20° (Broili 1967; Semenza 2001, 2010). Unfortunately, this idea of the (sub)horizontal 'seat' was further reinforced by a critical error by Semenza (1965, Section C.5). In July 1960, three boreholes were drilled with the aim of locating the layer of 'cataclastic' material observed in the 'eastern wall' of the south side of the gorge (see Section 2.2, (i), above) that had been interpreted as the failure surface of the ancient landslide. This material was not encountered in any borehole. Consequently, Semenza suggested that this layer must have extended 'almost horizontally' much further from the gorge, i.e. beyond borehole S1 (furthest from the gorge within the western half of the landslide: Fig. 2).

This interpretation was also partially based on an expectation that the drilling would encounter substantially intact rock of the Vaiont Limestone below the level of the failure surface, but no such 'solid' rock was encountered (Müller 1964). Boreholes S1 and S2 were located close to the most severely folded part of the failure surface (Fig. 2), i.e. in an area of high deformation resulting from



Fig. 7 Left wall of the Piave Valley above Longarone in September 2011, showing the ‘chair’ structure (bedding indicated by yellow lines) that was mistakenly assumed to extend beneath at least the western half of the landslide. The white line marks the alignment of the Col delle Tosatte Fault. The downstream end of the Vaiont Gorge is at the lower left of the image and part of the landslide scar is visible in the upper left part. Photo: APD

interference of two synclinal compression structures (Massironi et al. 2013). Whilst the Vajont Limestone appeared largely intact where exposed in the Vaiont gorge, in the light of this recent geological understanding it could perhaps be expected to be more highly fractured in this part of the slope and thus not easily differentiated from the overlying Fonzaso Formation. Thus was accidentally created a defining characteristic of the Vaiont landslide that most researchers have subsequently followed.

Shape of the concealed failure surface: not a ‘chair’ but a bowl

It is the geometry of the ‘chair’, with its steep ‘back’ and horizontal or subhorizontal ‘seat’, and/or failure to take account of the alignment of such cross-sections with respect to a 3-dimensional bowl-shaped failure surface (Müller 1964; Kiersch 1964; Broili 1967), that has confounded previous attempts to provide a single explanation of all aspects of the event. However, NE-SW tectonic compression deformed the ‘chair’ structure associated with the Erto syncline east of the N-S Col delle Tosette fault (Fig. 6). This ‘Stage 2’ compression (see Section 2.2, above) formed the Massalezza syncline and created a bowl-shaped 3D structure

(Massironi et al. 2013) which, when viewed in N-S subvertical cross-section in the left wall of the Piave valley opposite Longarone, happens to look like a ‘chair’ (Fig. 7).

Müller (1961, cited by Semenza 2001, 2010) had noted that beneath the eastern side of the landslide the basal failure surface was generally much steeper, as we surmised from our initial investigations (Dykes et al. 2013). Furthermore, Kiersch (1964) and Broili (1967) had suggested that the failure surface was ‘bowl-shaped’ in three dimensions, with Broili’s interpretation (Fig. 8a) being adopted by Müller (1968). The most likely form of the failure surface was identified very early as being broadly concave in plan and section eastwards from the western margin, stepping between several beds of the east-dipping Fonzaso Formation, but then steeply stepped upwards through the overlying strata towards the Col Tramontin Fault which largely determined the position of the eastern margin (Müller 1964, 1968). The latter configuration constituted a relatively small area of very high resistance to movement in the northeast corner of the landslide area that explains the pattern of recorded movements during 1960–63 (Müller 1964; Belloni and Stefani 1992; Paronuzzi and Bolla 2012;

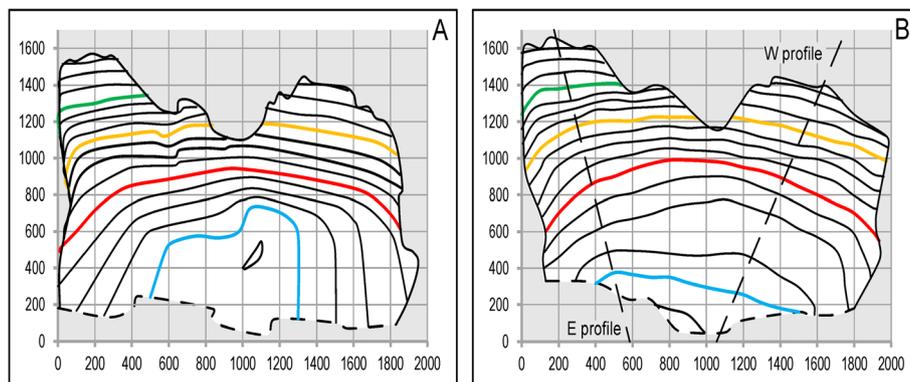


Fig. 8 a Map of the failure surface according to Broili (1967) as adopted by Müller (1968). b Map of the failure surface according to Bistacchi et al. (2013), showing the lines of the two profile cross-sections in Fig. 9. Both maps use the same arbitrary grid as Fig. 2 with axis scales in metres. The outline of the 1963 landslide is shown by the thick black line, with the outcrop of the failure surface along the gorge indicated by the broken black line. Contours are at 50 m vertical intervals increasing upslope away from the gorge: blue = 600 m, red = 800 m, dark yellow = 1000 m, green = 1200 m

Wolter et al. 2014). Despite these three-dimensional interpretations, most researchers analysed the stability of the Vaiont landslide in terms of approximately S-N cross-sections, each having a steep subplanar upper failure surface and a subhorizontal lower failure surface (e.g. Guidici and Semenza 1960; Müller 1964, 1968; Semenza 1965; Mencl 1966; Skempton 1966; Broili 1967; Romero and Molina 1974; Chowdhury 1978; Hendron and Patton 1985; Belloni and Stefani 1992; Hutchinson 1987; Tika and Hutchinson 1999; Kilburn and Petley 2003; Alonso et al. 2010; Alonso and Pinyol 2010; Pinyol and Alonso 2010; Hungr and Aaron 2013; Paronuzzi and Bolla 2012, 2015; Boon et al. 2014; Del Ventisette et al. 2015; Havaej et al. 2015; Paronuzzi et al. 2016; Wolter et al. 2016; Crosta et al. 2016).

We derived, from site observations in 2011 and 2012 and from original mapping (e.g. Rossi and Semenza 1965), an approximate estimated geometry for the failure surface, in order to explore hypothetical stability conditions by modelling. This estimated failure surface did not have a ‘chair’ structure. Then, at the 50th Anniversary conference, Bistacchi et al. (2013) presented a new 3D digital model that provided contour maps of the sliding surface shape quantitatively derived from comprehensive and verified geological data, including re-interpretations of borehole

information, using geostatistical interpolation techniques. These maps were very similar to our own attempts and likewise did not show the existence of a ‘chair’ shape, thus they unknowingly confirmed our initial proposition. We only had access to a subset of the information that Bistacchi et al. used, so we set aside our own assessment in favour of their more rigorous and comprehensive version. Consequently, it now appears that the failure surface of the 1963 landslide is indeed *not* ‘chair’-shaped and does not have a (sub)horizontal component (Figs. 8b and 9). This critical finding invalidates consensus point C₃ (Table 1).

The landslide of 4 November 1960—F4

Filling of the reservoir commenced in February 1960. A small rockfall of March 1960 occurred from the vertical wall of the gorge—well above the water level—near the upstream extent of the 1963 failure zone. This appears, from inspection of a photograph taken at the time (Fig. 19 in Semenza 2001, 2010), to have been an example of the regular natural geomorphological events that should be expected in this context. Around the same time in early 1960, a 350-m wide strip of rock along the northern edge of the Pian della Pozza, only a short distance upstream of the dam, was noted to be moving. This mass had subsided by around 10 m

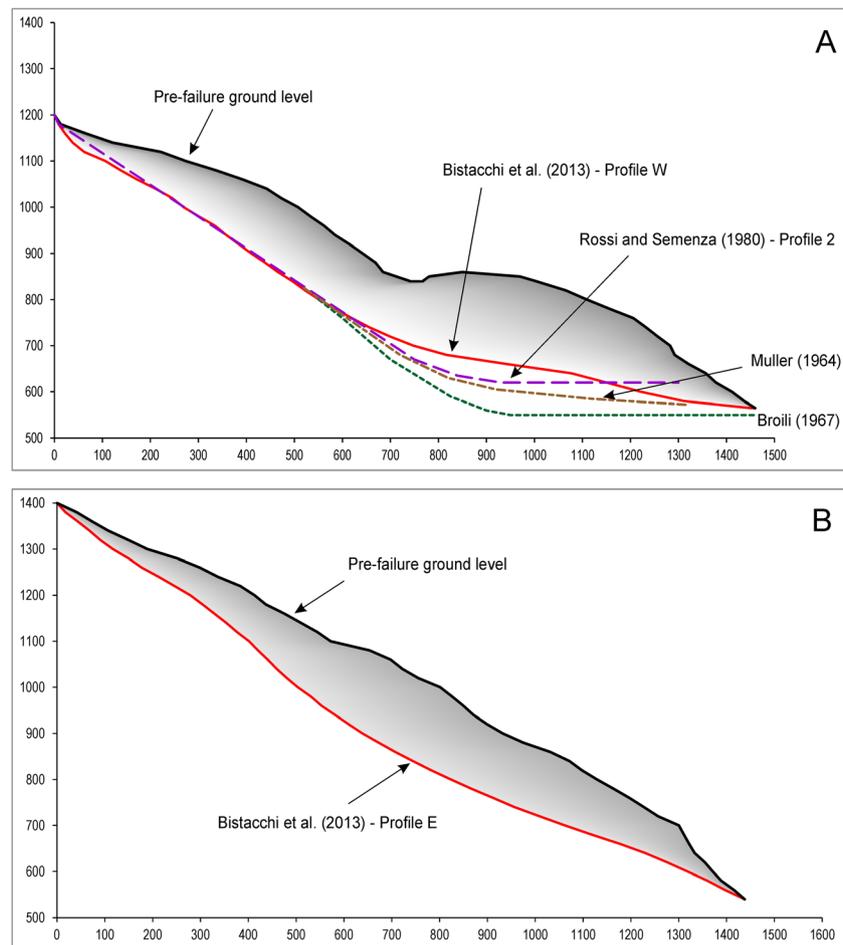


Fig. 9 a Cross-section through the western part of the Vaiont landslide along the line of Bistacchi et al.’s (2013) ‘Profile W’, also showing some previous failure surface interpretations superimposed on this profile. Note that the profile of Rossi and Semenza (1980) does not correspond at the toe due to its orientation being slightly different. b Bistacchi et al.’s ‘Profile E’. All axes in metres; the vertical axes are elevations above mean sea level

when photographed by Semenza in April 1960, and collapsed fully as a 700,000-m³ landslide on 4 November 1960 (Fig. 2). By this time, the reservoir water level had risen to almost 650 m (i.e. ~70 m above the reference elevation). Semenza (2001, 2010, caption to Fig. 35) wrote: ‘This small movement cannot be placed in direct relationship to the reservoir. At that time, in fact, the filling had not yet reached the lower limit of the failure surface of the slide of November 4. ... the movement of Nov. 4 was the final phase of that little slide ...’.

Cross-sections from the new failure surface geometry (e.g. Fig. 9a) show the failure surface becoming steeper towards the gorge in the west side of the slope, i.e. corresponding with photographs of the bedding below the western edge of the Punta del Toc at the eastern end of the November 1960 landslide (i.e. contrary to C3). Kiersch (1964) had identified stress relief joints parallel to the upper edge of the gorge, also visible in contemporary photographs (Masé et al. 2004). There was thus an inherent instability that was not triggered by the rising reservoir level (indeed, we have found no reference to this possibility) but was probably initiated by rainfall-induced raised groundwater pressures below the 1963 landslide failure surface (see ‘F5’ Section, below). The November 1960 rockslide unloaded a 350-m width of the toe of the west side, possibly resulting in new stress relief joints and probably reducing the net resistance to shearing.

Effect of reservoir levels and rainfall on pore water pressures—F5

Raised pore water pressures due to impoundment of the reservoir are widely regarded as the cause of the landslide (consensus point C4), whether as a first-time failure or a reactivation (Kiersch 1964; Semenza 1965; Chowdhury 1978; Hendron and Patton 1985; Hutchinson 1987; Alonso et al. 2010; Alonso and Pinyol 2010; Pinyol and Alonso 2010; Ward and Day 2011; Delle Rose 2012; Paronuzzi and Bolla 2012, 2015; Paronuzzi et al. 2016; Havaej et al. 2015; Wolter et al. 2016). Unusually for the period there are some measurements of water levels and ground movements from the southern side of the Vaiont gorge prior to the landslide (Müller 1964), although these are extremely limited.

The period of the first filling (February to November 1960) coincided with a very wet summer (June onwards) followed by an exceptionally wet autumn (around 425 mm of rain fell in October at Erto: Müller 1964) (Fig. 10). Significant movement of the northern slopes of Mt. Toc began in late September. Critically, it is often overlooked that during the first filling of the reservoir, the rate of movement decreased markedly during the first 4 days of November, even whilst the reservoir level was still rising, because it had stopped raining (Müller 1964; Hendron and Patton 1985). Semenza (2001, 2010, footnote 39) reported springs located ‘at the base of the mass’ (of the ancient landslide) with one in particular noted below the northwestern corner of the Pian del Toc, i.e. only a short distance from the eastern edge of the November 1960 landslide (Fig. 2). According to this observation which indicates significant macropore drainage (even if not correctly associated with the future failure surface position), it is also possible that the decrease of movement rate would have been even steeper if a 70-m depth of the lower gorge slope had not been inundated, because this may have impeded groundwater drainage from the toe of the landslide mass.

Rainfall fell to more normal levels as the reservoir level was reduced to allow the bypass tunnel to be built, and movement of the entire mountainside stopped. During the second filling of the reservoir (October 1961 to December 1962), rainfall was normal and measured ground movements were of the order of 1–2 mm day⁻¹. With the external water level significantly higher than the 650 m attained in November 1960, the unstable mass did not accelerate significantly during the second filling until another period of exceptional rainfall (> 300 mm during 20 days in late October/early November 1962), with prior small accelerations during April–July 1962 also correlating with somewhat higher rainfall. The small ongoing movements probably resulted from propagation of cracks formed during the first filling and, as such, may be expected to have continued irrespective of internal or external water conditions. The final acceleration in August 1963, that ultimately continued to the landslide, can also be seen to relate to higher rainfall, though not as high as the previous accelerations. All three of the major accelerations, i.e. in November 1960,

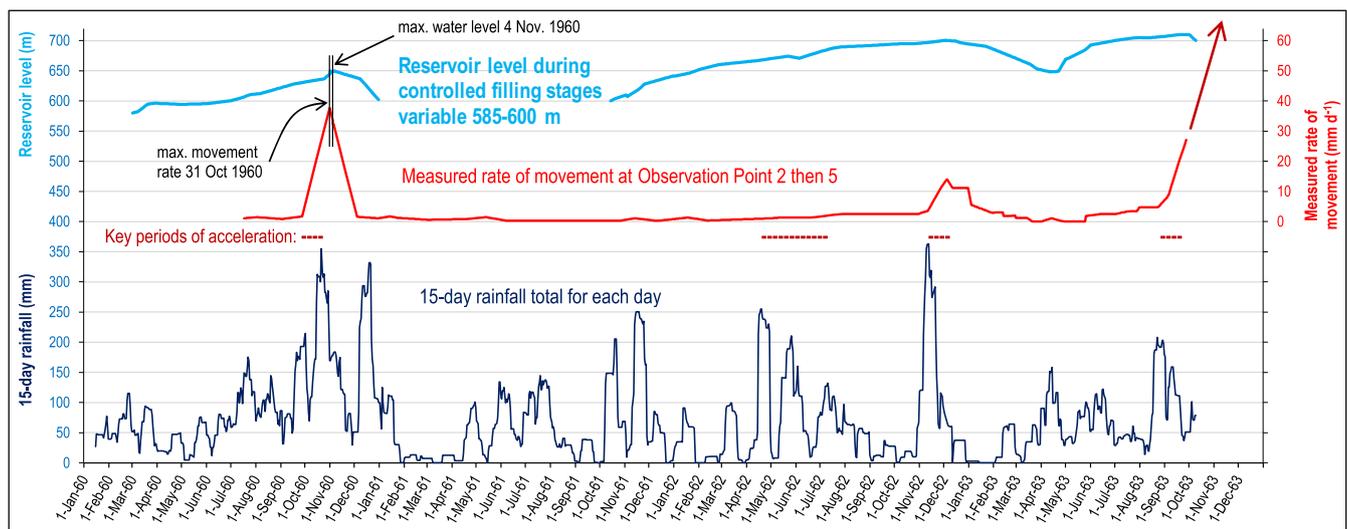


Fig. 10 Composite graph of 15-day rainfall totals at Erto (derived from data presented in Hendron and Patton 1985, Appendix A), reservoir water level elevation and movement rates in the lower western part of the developing landslide (Müller 1964), for the period 15 January 1960 to 9 October 1963 inclusive

December 1962 and August 1963, coincided with distinct phases of seismic activity (Delle Rose 2012) as examined in the ‘F6’ Section (below). The strong correlation of movement rates with rainfall, whilst not necessarily implying a causal relationship, nevertheless indicates that C4 may not be valid.

During September–October 1961, four piezometers were installed within the north side of the mountain slope (Fig. 2) although piezometer P4 ‘failed early’ and yielded no data (Müller 1964). Data from P2 initially showed water levels up to 90 m higher than the reservoir level, which was thought to indicate significantly artesian groundwater pressures (Müller 1964; Hendron and Patton 1985), but later it showed a water table consistent with P1, P3 and the reservoir elevation.

Replotting these data with the failure surface elevations of Bistacchi et al. (2013) at the piezometer locations (Fig. 11) reveal details contrary to any previous account. The failure occurred along clay layers at varying stratigraphic positions within the Fonzaso Formation, so in reality ‘failure surface elevation’ means an indicative position that may vary by several metres higher or lower than the stated value.

Previously, because of the assumed ‘chair’ structure (consensus point C3), it was thought that the bases of P1 and P3 at around 620 m elevation (Semenza 1965), and the observed water levels in P2, must have been located above the failure surface within the landslide mass (Hendron and Patton 1985). The P2 data were interpreted as showing significant artesian water pressures acting on the base of the landslide mass, an effect which must (under C3) have been so much greater beneath a large subplanar failure surface at around 600 m elevation. It now seems that there is no evidence of artesian groundwater beneath the failure surface (Figs. 11 and 12), although there is a small possibility of such conditions near the upstream end of the toe of the slope where the clay layers descend below the floor of the gorge. Most of the landslide mass was therefore underdrained, although the clay layers probably significantly restricted any vertical drainage. It now seems that P4 may have been seated in the unsaturated zone below the clay layers but above the lower aquifer, and was abandoned before being affected by the rising reservoir-induced groundwater.

The much higher initial water levels in P2 compared with P1 can now be seen to result from a ‘lower aquifer’ water table, at least 20 m below the failure surface elevation, that probably broadly follows the geometry of the bedding (Fig. 12). P3 was affected by the reservoir first because it was located closer to the gorge than P1; P2 was much further away from the gorge and higher up the inclined bedding; hence, it remained unaffected for longer. Reports of solution cavities by Hendron and Patton (1985) suggested a relatively high overall permeability for the Vajont Limestone below the landslide mass; responses of P1 and P2 to rainfall may have supported this observation but conversely may have been influenced by piezometer leakage from above the clay layers (see below).

Given the apparent absence of artesian water pressures, there remains the issue of landslide movements being directly correlated with rainfall and not with reservoir water levels. The only plausible explanation for this is that a transient ‘perched’ water table formed above the clay layers within the landslide mass following rainfall, i.e. Semenza’s (2001, 2010) ‘upper aquifer’. There are no data that show groundwater conditions within the landslide mass, but our new interpretations provide a framework for future modelling investigations of realistic possibilities. However, the fact that the piezometers were unsealed (Hendron and Patton 1985) allows some inferences to be drawn. They passed through—and thus allowed hydraulic connectivity across—the clay layers, yet showed piezometric levels significantly below the landslide mass. This suggests that the landslide mass was largely dry except briefly following periods of heavy rainfall and inundation by reservoir water. This explains the piezometer water levels aligning with reservoir levels and then tracking them as hydraulic continuity became established through the unsealed piezometers.

We subscribe to the detailed analysis of Hendron and Patton (1985) that appears to demonstrate a very strong rainfall component which, when superimposed on the local reservoir level effect, explains the observed patterns of movement of the slope. This partially invalidates consensus position C4. Furthermore, although Hendron and Patton’s analyses incorporated C1, C2 and C3 (Table 1), they derived thresholds for

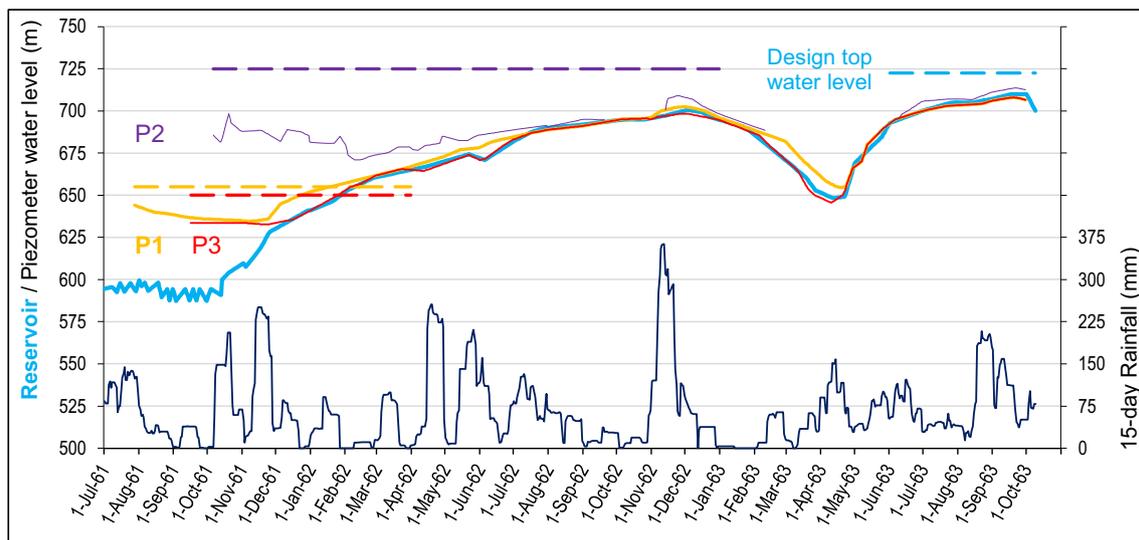


Fig. 11 Composite graph of reservoir and piezometer water level elevations in the lower western part of the developing landslide (Müller 1964) and 15-day rainfall totals at Erto (derived from data presented in Hendron and Patton 1985, Appendix A), for the period 1 July 1961 to 9 October 1963 inclusive. The elevation of the failure surface at each piezometer is indicated by the broken line in the same colour

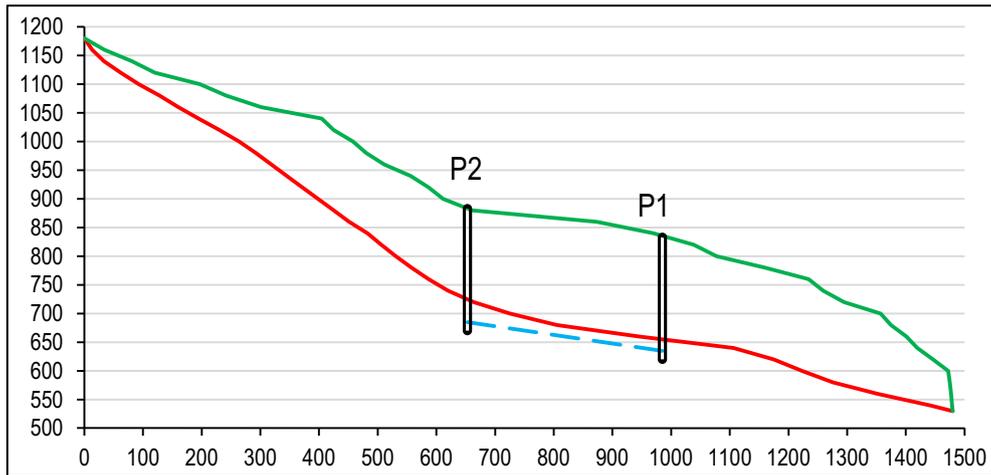


Fig. 12 Cross-section through the west side of the landslide aligned with piezometers P1 and P2 showing the recorded water levels around November 1961 (Müller 1964) and an indicative 'lower aquifer' water table (broken blue line) between them. Based on Semenza (1965), both west side piezometers are assumed to be 220 m deep as shown here. Both axes in metres; the vertical axis indicates elevation above mean sea level

instability based on combinations of reservoir water levels and multi-day rainfall totals, particularly the 15-day rainfall totals (also used to define landslide thresholds by Lumb (1975), Pasuto and Silvano (1998) and Bell and Maud (2000)) shown in Fig. 10, and concluded (i) that the landslide would have occurred without rainfall if the reservoir had reached its design level of 722.5 m, and (ii) that without the reservoir, 7-day rainfall of 180 mm or 15-day rainfall of 350 mm—exceeded twice in Fig. 10—would be sufficient to cause instability. Similar analyses now need to be redone, using the new first-time failure surface geometry with peak shear strength

values, to determine how close to collapse were the slopes of Mt. Toc even without the construction and impounding of the reservoir.

Eastward development of the failure surface—F6

Ground movements were measured at the two observation points referred to above and at up to 22 further positions across the northern slope of Mt. Toc (Table 1 in Müller 1964). Measurements show that during the 'first filling' of the reservoir during 1960, the largest velocities were recorded in the lower west side of the slope,

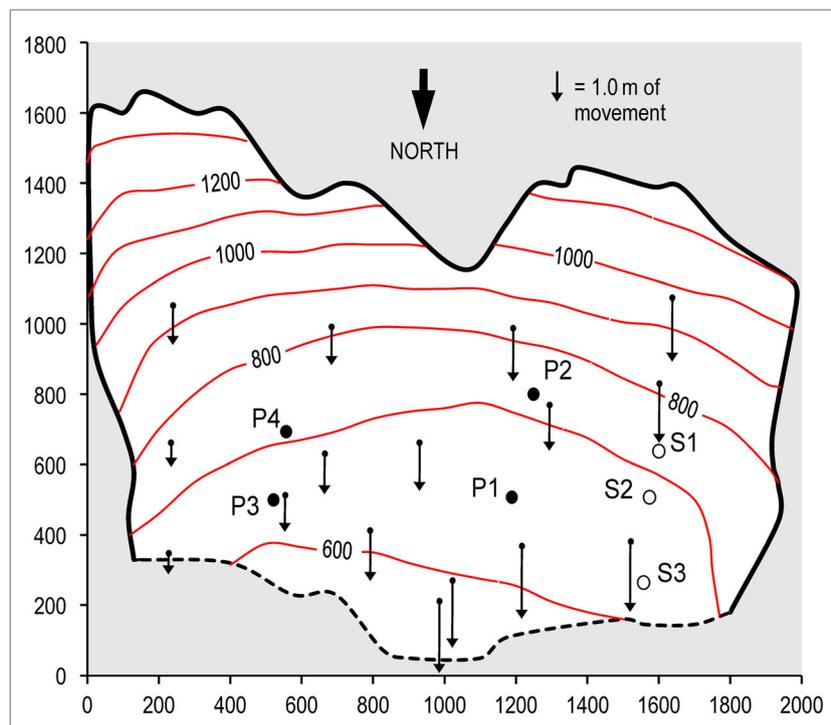


Fig. 13 Simplified map of the failure surface of the landslide (after Bistacchi et al. 2013) showing the locations of boreholes S1–S3, piezometers P1–P4 and total displacement magnitudes at 16 monitoring sites (not vectors; arrow lengths scale with displacement) between December 1960 and the end of September 1963 (Müller 1964). Axes and contour values are in metres

with movements above the Pozza being lower by about half. Velocities were lower still in the upper east side, reducing towards the gorge east of the Massalezza Ditch (Fig. 13). This pattern is consistent with a down-dip (i.e. ENE) vector of shear stress on the failure surface modifying the downslope trend, in which the lower west side is entirely unsupported but the lower east side is supported and resisted by the great thickness of intact beds stratigraphically above the failure surface.

In addition to the above, seismic investigations were undertaken by Professor P. Caloi in November 1959 and December 1960 within the landslide, and a seismograph was installed on the Vaiont dam for the duration of the reservoir filling. Belloni and Stefani (1992) examined Caloi's geophysical data in detail and highlighted the initial seismic results as indicating essentially 'in situ rock' in 1959 (Caloi 1966) with a deterioration in the quality of the rock mass between 1959 and 1960 (Havaej et al. 2015) due to 'fracturing processes in the deeper lying rock layers' (Müller 1964, p.168). Semenza (2001, 2010) argued that geologically this was inconsistent with observations of only 1 m of movement of the otherwise in situ west side of the mountain during 1960, implying that much greater movement should be needed to produce such a degree of rock mass deterioration, but the published accounts of Caloi's studies and the interpretations of his data appear sound. The December 1960 data indicated 'much fractured rock' down to around 640 m elevation near the gorge (Müller 1964, p. 159; Alonso and Pinyol 2010). Delle Rose (2012) later reported Caloi's (1966) interpretation of seismicity within the northern slope of Mt. Toc generally migrating eastwards during the 1960–63 monitoring period, consistent with the observed pattern of movement.

These observations are inconsistent with consensus positions C1 and C2. If there was an ancient failure surface at residual strength, the measured displacements indicating slow slip would most likely not have produced detectable seismic signals because

there would have been no cracking of intact rock. In fact no seismic signals from slow slip events are known to have been recorded up to 5 years ago (Tonnellier et al. 2013). Furthermore, if there was an entire landslide at residual strength, it is difficult to imagine a mechanism that would cause it to reactivate sequentially from one side towards the other.

The movement of the landslide—F7

The primary problem for previous investigations of the Vaiont landslide has been to explain how the failed mass was able to accelerate to an extraordinarily high velocity and, as a result, achieve such a large displacement, given the assumed 'chair'-shaped failure surface (C1). Whilst the magnitude of the displacement is not in doubt, having been recently re-analysed using modern GIS techniques and found to be around 360 m for the western half and around 460 m for the eastern half (including 100 m over the displaced western half) (Bistacchi et al. 2013), there remains some debate regarding the duration and maximum speed of the movement. Table 4 summarises the variations in estimates based on different factors including kinematics (e.g. Müller 1964), seismic analysis (Caloi 1966; Della Rosa 2012) and displacement wave simulations (Ward and Day 2011; Crosta et al. 2016; Zhao et al. 2016). This latter approach appears to be generating results that are converging with the more extreme (rapid) range of previous estimates. However, the model results must show the landslide movement into the reservoir to have a velocity profile sufficient to achieve the observed wave limits. Ward and Day's (2011) results required an east-to-west development of the landslide with movement initiated within the eastern side of the landslide in order to make the simulated flood correspond with the observed flood limits, i.e. it is inconsistent with other analyses of the landslide, and Crosta et al.'s (2016) results rely on the subhorizontal 'seat' of the 'chair' structure being at residual strength ($\phi_r = 6^\circ$, i.e. assuming C1–C3). Zhao et al. (2016) used the new failure surface geometry (Fig. 8b). Taking the 'worst-case' estimates in Table 4, we will suggest

Table 4 Estimates of the movement of the landslide

| Maximum velocity (m s^{-1}) | Duration of movement (s) | Reference |
|--|--|--|
| 30 | 30 | Kiersch (1964) |
| 25–30 | 30–45 | Müller (1964) |
| | 20–25 | Caloi (1966) |
| 25–30 | 50 | Crosta et al. (2016) |
| 20–30 | < 60 | Romero and Molina (1974) |
| 20–30 | 'Few tens of seconds' | Hendron and Patton (1985, p.8) |
| 30 | 45 | Kilburn and Petley (2003) |
| | < 45 | Genevois and Ghirotti (2005) |
| ~ 25 | 45—east side | Zhao et al. (2016) |
| ~ 22 | 50—west side | |
| 17 | 45 | Selli et al. (1964), cited in Genevois and Ghirotti (2005) |
| 17 | 45 | Ciabatti (1964), cited in Genevois and Ghirotti (2005) |
| 15 | | Nonveiller (1978), cited in Genevois and Ghirotti (2005) |
| 15–20 | ~ 45–50 (finished within 2 min) | Ward and Day (2011) |
| | 120 | Delle Rosa (2012) |
| | 45–50 (causing seismic shocks lasting 100 s) | Selli and Trevisan (1964), cited in Delle Rosa (2012) |

an explanation that can account for 400 m displacement in 30 s, simplified as a uniform acceleration of 2 m s^{-2} increasing the velocity from 0 to 30 m s^{-1} over 15 s, followed by a uniform deceleration at the same rate. This explanation will also account for any lower acceleration and velocity if the landslide did move somewhat less rapidly in reality.

Retrogression of the landslide head—F8

The east side of the landslide was almost translational in nature (Fig. 9b). The upper part of the east slope did not initially fail, but moved later—after the destructive waves had diminished—and came to rest on top of the main failed mass with most of the tree cover still standing. This is the ‘eastern lobe’ of Semenza (2001, 2010) and others. There is one colour photograph in the National Park Visitor Centre in Erto village (which is dedicated to the Vaiont disaster), taken immediately after the landslide in daylight the next morning (10 October 1963), that shows a similar but smaller part of the upper west side that behaved similarly. Wolter et al. (2016) identified the retrogressive blocks from both sides of the landslide as ‘blocks C and D’ in their recent study. Hence, there is contemporary photographic and modern analytical evidence of retrogressive development of both sides of the landslide after the flood had subsided.

Crosta et al. (2016) determined from simulations of the reservoir displacement wave that the total period of significant water motion was around 144 s (2D analysis), whereas Zhao et al. (2016) obtained results showing the wave velocities and water levels to have stabilised within around 60 s. Therefore, the ‘eastern lobe’ must have failed at least a minute after the main landslide movement had occurred, with the western side retrogression taking place many hours later. Hendron and Patton (1985) disregarded the ‘eastern lobe’ as not necessary for an understanding of the main slide. However, retrogression of the head immediately post-failure, within the timescales indicated, demonstrates a smaller disturbing force on the steep upper part of the failure surface at the moment of release of the main landslide mass, making the possibility of failure on a ‘chair’ shaped structure (C₃) even less likely.

Examining the existing consensus position

Internal consistency: clay seams, pre-existing slide, ‘chair’ shape

Most previous studies of the Vaiont landslide have used cross-sections through the western side of the landslide for both limit equilibrium and kinematic analyses of stability conditions. The reasons for this arose from (i) the focus of the investigations on the areas nearest the dam due to initial concerns about the safety of the dam; (ii) observations that the western part was moving so much faster than the eastern part (Müller 1964; Fig. 11); and, after the landslide, (iii) the general belief at the time that the western side had the greater extent of subhorizontal shear surface (C₃) and that understanding this part of the landslide was essential in order to explain the whole event. Hendron and Patton (1985) presented six representative cross-sections through the west side of the landslide, from previous studies, together for comparison (similar to Fig. 9a). All are similar, but none are identical and, necessarily given the lack of relevant data, other subsequent cross-sections have been similarly variable. In connection with (iii) above, Rossi and Semenza (1980) prepared three cross-sections for Hendron and Patton to use for their investigation. Identified as Sections 2, 5

(both in the western side) and 10A (a short distance east of the Massalezza Ditch), they were ‘selected as representative sections for use in stability analyses. They were also chosen because they appeared to be oriented relatively close to the direction of the original movement of the slide’ (Hendron and Patton 1986, p. 85). Subsequent studies have used these sections (e.g. Alonso and Pinyol 2010) or others similar to Section 2 in particular (e.g. Paronuzzi and Bolla 2012), i.e. further reinforcing consensus point C₃.

In line with the above, many studies of the landslide utilised 2D limit equilibrium stability analyses to determine the shear strength that must have operated on a failure surface with some form of ‘chair’ shaped geometry. If the 1963 failure was a reactivation of an ancient landslide then the failure surface must have been pre-sheared and therefore must have existed before October 1963 at residual strength. Moreover, if the slide had taken place along clay seams, then that residual shear strength should be the residual shear strength of a clay, in which case a back-analysis of the stability of the slope (e.g. Chandler 1977) as it existed prior to October 1963 would show this. Indeed, angles of internal friction obtained from these analyses, all described as representing the residual shear strength of the clay layers, generally fall in the range $6\text{--}12^\circ$ (Semenza 2001, 2010; Hendron and Patton 1985, 1986; Hutchinson and Kwan—unpublished manuscript 1986; Alonso and Pinyol 2010). The widely accepted assumption of a ‘chair’ shaped slip surface is seductive, because using the ‘chair’ geometry yields a low mobilised strength for the pre-movement conditions—corresponding with residual strength of clay—and this simply confirms the ancient landslide hypothesis in the mind of the investigator. Thus, reactivation of an ancient landslide with a ‘chair’ shaped failure surface located along thin clay beds at or near residual strength constitutes an internally consistent hypothesis, hence consensus element C₂ (Table 1). Our new analyses using the Bistacchi et al. failure surface shape (Dykes and Bromhead 2018) demonstrate shear strengths prior to the collapse of the slope that are consistent with unsheared clays and not with sheared clays, providing further evidence contrary to C₂ and therefore also C₁.

Influence of ‘chair’ shape on assumed trigger mechanism (impoundment)

Figure 9a shows cross-sections through the landslide by Broili (1967) and Rossi and Semenza (1980) that have horizontal lower segments and very steep upper parts. Many other assumed cross-sections similarly have (sub)horizontal lower segments. If there was such a large extent of almost horizontal basal shear surface (C₃), any increase in reservoir surface elevation above the level of the basal shear surface should have raised pore water pressures across a significant proportion of the entire failure surface (e.g. up to 1 MPa by November 1960), thus reducing the effective shear strength accordingly. At the same time, the beneficial effect of the reservoir, i.e. providing lateral support to the toe of the slope, was very limited because the submerged depth of the future landslide mass above the basal shear surface was little more than the same 100 m that produced the pore water pressures.

We have not found a published study of Vaiont that does not attribute the initiation of movement in 1960 to the first filling of the reservoir. This appears reasonable given the correspondence of movement initiation with submergence, at least during the first

filling, given the implications of C₃ outlined above. Indeed, this argument further supports the consensus (C₄ in Table 1) in providing a mechanism for a significant (further) reduction in effective shear strength below the assumed residual values. However, the rather more translational and inclined failure surface (Fig. 9) significantly reduces the reservoir effects because of the lower pore pressures at the failure surface and the smaller spatial extent of failure surface affected by these raised pore pressures. Together with the observed correspondence between rainfall and accelerations of movement, these factors provide further grounds to reconsider C₄.

The effect of pore pressures arising from deep circulation of unusually high rainfall may have been systematically overlooked in the past but it now appears to have been less relevant than some reports indicated (Section 2.3, above). The moderate dip of the bedding towards the gorge throughout the lower half of the landslide mass promoted free drainage of infiltrated rainwater rather than the development of artesian pressures as previously thought. The most critical element of the problem is that element for which there are no data, i.e. groundwater conditions within the landslide mass that seem to have dominated the initiation and subsequent accelerations of the movement of the mass, particularly upslope of the zone of reservoir influence (Fig. 10).

Issues with the consensus

An ancient landslide with a ‘chair’ shaped failure surface at residual strength (C₁–C₃ in Table 1) could not have accelerated and achieved the observed speed and runout without some other unusual mechanism having arisen within the slope to account for this behaviour (c.f. Hutchinson 1987). Increasingly complex mechanisms have been proposed over the years to explain further reductions in shear strength below the residual value that could account for the observed acceleration, assuming the consensus position to define the problem. For example, Romero and Molina (1974) concluded that the residual strength in the reactivated slide was probably between 14° and 18° with heating effects overcoming some shear resistance during failure. Frictional heating was also advocated by Voight and Faust (1982), Anderson (1985), Nonveiller (1992) and Pinyol and Alonso (2010). Other hypotheses include the loss of internal rock strength associated with internal shears (e.g. Hutchinson 1987; Alonso and Pinyol 2010), which the latter authors acknowledged would have resulted in further loss of (kinetic) energy thus leaving unanswered questions, or brittle failure of shear surface asperities and/or shear surface gouge that had recovered some brittleness post-failure (Hutchinson 1987).

Those authors whose starting point was that the 1963 landslide was a first-time failure (but, following the consensus C₃, with a ‘chair’ shaped failure surface along thin clay beds) had to explain how the stable mountain slope at peak shear strength could overcome the restraining effect of the subhorizontal lower part of the assumed failure surface and start to move, i.e. how was enough strength lost. Some authors suggested that there may have been reduced shear strengths due to tectonic shear or flexural slip along bedding planes between limestone beds (e.g. Dedic 1987; Delle Rose 2012). Hendron and Patton (1985) also highlighted this possibility although some of this strength may have subsequently been recovered, as suggested by Hutchinson (1987) with respect to a hypothesised ancient landslide at the site. Such tectonic shearing

could explain the occurrence of residual strength without needing an ancient landslide, but this would contradict their (Hendron and Patton 1985) primary explanation. The possible role of flexural slip as a significant contributor is in any case greatly reduced by the much less severely folded geometry of the new failure surface.

By contrast, Müller (1964, p. 199) wrote that the landslide was: ‘... probably explained by a sudden and considerable – although unknown and unrecognisable – decrease of the internal friction of the slide mass’, with a final loss of strength due to ‘some parts of the base ... [being] inexplicably sheared off’ (p.209). He later wrote in terms of what we now refer to as ‘progressive failure’ (Müller 1968), also later suggested by Belloni and Stefani (1992); Nonveiller (1992) invoked such a mechanism as part of his ‘reactivation’ explanation. More recent studies have suggested plausible mechanisms for first-time failure—Kilburn and Petley (2003) showed experimentally that brittle progressive failure by crack propagation could occur within the clay interbeds at the range of stresses applicable to the Vaiont context and Havaej et al. (2015) demonstrated a plausible 3D simulation of brittle rock fracture—though still not accounting for both the occurrence *and* behaviour of the slide.

It is now clear that some of the earliest accounts of the Vaiont landslide presented and discussed details and interpretations that were very close to those of some very recent investigations. Broilli’s (1967) failure surface is, in principle, very close to Bistacchi et al.’s (2013) version, although too extreme in its curvature (Fig. 8). Müller’s (1964, 1968) assessment that there was no ancient landslide and that the available evidence indicated progressive failure developing through an essentially intact rock mass appears to be more realistic than the consensus, although his suggested ‘glacier’ mechanism has not been supported by any other study. Kiersch (1964) sketched a cross-section that appeared more similar to Fig. 9a and highlighted the probable role of high rainfall contributions to groundwater flow via karstic voids in the limestones. It is perhaps unfortunate that Semenza’s accumulation of significant evidence apparently consistent with his primary hypothesis of an ancient landslide, thereby establishing consensus points C₁, C₃ and C₄ (with C₁ leading to C₂), gained more prominence and somehow became the default primary source for subsequent investigations.

Conclusion: a new hypothesis

Our investigation developed from an initial observation that the eastern part of the landslide could not possibly manifest the ‘chair’ shape assumed by the consensus (C₃). This led us to attempt a reconstruction of the shape of the entire failure surface, work that was rendered obsolete by the publication of a new map of the failure surface geometry by Bistacchi et al. (2013). Sections drawn through the landslide using the Bistacchi failure surface do not show a subhorizontal ‘chair seat’ even in the western part of the landslide and without that, some of the apparent consistency in the consensus model disappears. Moreover, having discovered that much of the current consensus position is based largely on one early misinterpretation (“Shape of the concealed failure surface: the ‘chair’” Section, above), then we have felt unconstrained to believe other aspects of the consensus and have re-examined all possible elements of the problem. We have always agreed with part of C₂ (Table 1), i.e. that the failure was founded on layers of clay within the limestone sequence—but at peak strength prior to 1960. However, we have identified several aspects of the 1963 landslide that cannot be reconciled with the remainder of the consensus position.

The logical argument is this: if the 1963 failure was a reactivation of an ancient landslide then it must have occurred at residual strength, in which case a back-analysis of the stability of the slope using the ‘chair’ geometry simply confirms this hypothesis in the mind of the investigator. Furthermore, an ancient landslide at residual strength could not have accelerated and achieved the observed speed and runout without some other unusual mechanism having arisen within the slope to account for this behaviour. However, if the 1963 landslide was a first-time failure then the reduction of shear strength from peak to residual is sufficient to produce the conditions required to account for the rapid large displacement of the unstable mass—but the ‘chair’ inhibits this possibility. If there is no (or very little) ‘chair’, there cannot have been an ancient landslide because the mass could not have remained in place at residual strength, and the role of rainfall becomes highly significant as a causal factor. Therefore, we arrive at a new hypothesis for the Vaiont landslide that largely rejects the consensus, i.e.:

- H1 The 1963 landslide was a first-time failure.
- H2 Failure took place along thin clay seams that were initially at peak strength.
- H3 The sliding surface did not have a ‘chair’ shape but was moderately translational.
- H4 The slope was geologically predisposed to fail but rainfall was the main preparatory factor. Inundation of the toe was the eventual trigger but this was probably insufficient on its own, only being effective in concert with the heavy rainfall.

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