Deposit temperature of pyroclastic density currents emplaced during the El Chichón 1982 and Colima 1913 eruptions

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Abstract: New data on the pyroclastic density current (PDC) deposit temperature (Tdep) are provided for two prominent eruptions of Mexican volcanoes of the twentieth century: the 1982 eruption of El Chichón and the 1913 eruption of Colima. In spite of similar lithofacies, magma composition and pre-eruptive conditions, the Tdep of the PDCs from the 1982 (El Chichón) and 1913 (Colima) eruptions differ significantly, with intervals of Tdep of 360–420 °C and 250–330 °C, respectively. These new data emphasize that a full understanding of the physical mechanisms responsible for equilibrium temperature attainment within a pyroclastic deposit has not yet been realized. The Tdep measured for El Chichón PDC deposits confirm the preliminary data published elsewhere, while Colima magnetic temperatures provide different values to those published previously.

Supplementary material: Tdep measurements for the different sites at El Chichon volcano and Colima volcano are available at: http://www.geolsoc.org.uk/SUP18695.

Pyroclastic density currents (PDCs) are moving mixtures of particles and gas that flow along the ground. They originate in different ways and from various sources during explosive eruptions or the gravity-driven collapse of domes (e.g. Druitt 1998; Freundt & Bursik 1998; Branney & Kokelaar 2002; Burgisser & Bergantz 2002; Sulpizio & Dellino 2008). They are among the most complex and dangerous volcanic phenomena, and pose serious problems for both human health and infrastructure in areas of tens–hundreds of square kilometres around volcanoes (e.g. Baxter et al. 2005; Gurioli et al. 2010). The direct measurement of physical parameters within moving PDCs is almost impossible, which highlights the need to obtain information from field and laboratory studies indirectly (e.g. Saucedo et al. 2004; Lube et al. 2007; Dellino et al. 2010; Girolami et al. 2010; Sulpizio et al. 2010; Sarocchi et al. 2011; Roche 2012).

Of particular interest is the assessment of the deposit temperature (Tdep) of PDCs, because it has relevance for both PDC dynamics and volcanic hazard mitigation. The Tdep of PDC deposits is a function of the initial temperature of the pyroclastic mixture, the size and provenance of its constituents, the cooling during transport and the sedimentation rate. Estimation of the Tdep can be addressed by measuring the thermal remanent magnetization (TRM) of lithic fragments in pyroclastic deposits. Lava fragments eroded from conduit walls or entrapped during transport carry, in principle, two TRM components. The high-T component is a record of the primary TRM acquired when the parent rock cooled down. Its direction in the clasts is random because the clasts were entrained from the parent rock, moved chaotically within the PDC and finally deposited in a randomly oriented position. The low-T component records the heating of...
the fragments and the eventual cooling during the PDC transport and in the deposit. The remanence of the magnetic grains in the lava clasts is characterized by blocking temperatures \( T_b \) which may vary from the ambient temperature up to the Curie point, \( c. 580 \, ^\circ C \) in the case of magnetite. The part of the primary TRM carried by the grains with \( T_b > T_{\text{dep}} \) is unaffected by the reheating of the clast within the current or the deposit, whereas that associated with \( T_b < T_{\text{dep}} \) is overprinted and results in a secondary TRM whose direction is parallel to that of Earth’s field at the time and the geographical location of the eruption. The threshold between the \( T_b \) spectra of the two components provides the reheating temperature, which is commonly interpreted to be the temperature of deposition \( (T_{\text{dep}}) \).

The palaeomagnetic estimation of \( T_{\text{dep}} \) has been recently investigated in different types of deposits of worldwide volcanoes: Somma–Vesuvius (Cioni et al. 2004; Zanella et al. 2007, 2008, in press; Di Vito et al. 2009); Stromboli (Porreca et al. 2006); Colli Albani (Porreca et al. 2008); Santorini (Bardot 2000; Bardot & McClelland 2000); El Chichón (Sulpizio et al. 2008); Taupo (McClelland et al. 2004); Unzen (Tanaka et al. 2004); and Cerro Galán (Lesti et al. 2011) among others. The interpretation of the results in these papers complies with the above model, with minor differences in the criteria used to estimate \( T_{\text{dep}} \).

Here we report the \( T_{\text{dep}} \) measured on different PDC deposits of the 1982 eruption of El Chichón volcano (Fig. 1a) (Carey & Sigurdsson 1986; Scolamacchia & Macías 2005) and of the 1913 eruption of Fuego de Colima volcano (Fig. 1b; Saucedo et al. 2010). Apart from a preliminary study on the 1982 El Chichón deposits (Sulpizio et al. 2008), previous studies of \( T_{\text{dep}} \) were performed on lahar deposits of Fuego de Colima (Paterson et al. 2010) and a debris avalanche of Nevado de Colima (Clement et al. 1993). The present data are the first comprehensive TRM study of PDC deposits from Mexican volcanoes and provide valuable information on \( T_{\text{dep}} \) from two of the most important explosive eruptions of the past century in Mexico.

**Volcanological setting and sampling**

El Chichón is a Late Pleistocene–Holocene volcano located in the state of Chiapas in southern Mexico. Prior to the 1982 eruption, El Chichón consisted of a 2-km-wide Somma crater, a central trachyandesitic dome and peripheral domes. In 1930, the El Chichón volcano had an episode of seismic unrest accompanied by fumarolic activity that lasted a few months (Müllerried 1933). Afterwards, El Chichón activity returned to background levels with the presence of weak fumaroles until 1980. New signs of volcanic unrest appeared in 1981 with increasing earthquakes, internal explosions and fumarolic activity reported by geologists and local villagers (Canul & Rocha 1981). The volcano reawakened in 1982, after 550 years of quiescence (Espíndola et al. 2000; Macías et al. 2003). The 1982 eruption began on 28 March with a 27-km-high Plinian column that deposited fall layer A1 (Fig. 2a; Carey & Sigurdsson 1986) and opened a 300-m-wide crater in the central dome (Medina-Martínez 1982). Afterwards, the volcano remained mostly calm with variable but impending seismicity (Jiménez et al. 1999; Espíndola et al. 2000). Late on 3 April, the eruption reawakened (Yokoyama et al. 1992) with a series of phreatomagmatic explosions that dispersed highly turbulent PDCs (PS1; Fig. 2a, b) around the volcano and killed more than 2000 people (Sigurdsson et al. 1984; Macías et al. 1997; Scolamacchia et al. 2005). These explosions opened the volcanic conduit and completely destroyed the central dome, producing block-and-ash flows (PF1, BAF1; Fig. 2a) that filled valleys around the crater. A series of small explosions emplaced PDCs (IU) in proximal areas followed by a 31-km-high column that deposited fall layer B (Fig. 2a). The column collapsed, generating pumice-rich PDCs (PF2; Fig. 2a) that further filled valleys around the crater and reached distances of 8 km along some valleys. This flow was followed by a dilute, turbulent PDC (PS2; Fig. 2a) that dispersed in all directions. After 4 h, the eruption restarted with another 29-km-high Plinian column that deposited fall layer C. Afterwards, phreatomagmatic explosions dispersed a dilute, turbulent PDC (PS3; Fig. 2a) that travelled up to 3 km from the vent. Subsequent activity produced minor dilute PDCs localized inside the new 1-km-wide crater reopened by the eruption. The eruption completely flattened c. 100 km² of jungle with the emission of 1.1 km³ of magma (Carey & Sigurdsson 1986) of trachyandesitic composition that had a pre-eruptive temperature estimated at 750–850 °C (Rye et al. 1984), 785 ± 23 °C (Luhr et al. 1984) and 800 °C calculated experimentally (Luhr 1990). On 8 April, temperatures in the PDC deposits measured with a thermocouple at 40 cm depth averaged 360 °C and were as high as 402 °C (Sulpizio et al. 2008).

**Fig. 1.** Location maps of El Chichón (upper) and Colima (lower) sampling sites. The inset in the upper right corner shows the position of the two volcanoes with respect to the Trans Mexican Volcanic Belt (in red; see colour in online version).
Five PDC deposits of the 1982 eruption of El Chichón volcano were sampled at 12 sites (Fig. 1a), from both massive (PF1 and PF2) and stratified lithofacies (PS1, PS2 and PS3).

Volcán de Colima (3860 m a.s.l.) is located within the western portion of the Trans-Mexican Volcanic Belt (Fig. 1) and constitutes the southern part of the Colima Volcanic Complex. It is the most active volcano in Mexico, and has produced c. 50 eruptions since 1576 (Medina-Martínez 1983; Luhr & Carmichael 1990; De la Cruz-Reyna 1993; Komorowski et al. 1997; Saucedo-Girón & Macías-Vázquez 1999; Bretón González et al. 2002; Saucedo et al. 2005; Cortés et al. 2010). Its historic activity includes Merapi- and Soufriere-type, subplinian and Plinian eruptions, as well as repeated gravitational collapses of the volcano edifice that produced large debris avalanches (e.g. Luhr & Prestegaard 1988; Stoopes & Sheridan 1992; Komorowski et al. 1997; Capra & Macías 2002; Cortés et al. 2005, 2010; Saucedo et al. 2005, 2010; Gavilanes-Ruiz et al. 2009; Roverato et al. 2011).

During this historic activity, Volcán de Colima has experienced at least 3 Plinian eruptions (in 1576, 1818 and 1913), 12 Soufrière-type eruptions and at least 9 Merapi-type events (Saucedo et al. 2005). The last large explosive event occurred during 17–20 January 1913. It occurred in three main phases (Fig. 2c, d): (1) an opening phase with the generation of Merapi-type pyroclastic flows (units F1, F2, F3) and a pyroclastic surge (S1; Fig. 2c); (2) a vent-clearing phase with strong explosions that produced Vulcanian-Soufriere-type pyroclastic flows (F4) and a pyroclastic surge (S2; Fig. 2c) that destroyed the summit dome, decompressing the magma system; and (3) a Plinian phase with the establishment of a c. 23-km-high column dispersing a fallout (C1) to the NE followed by the collapse of the column, which generated a pyroclastic surge (S3) and 15-km-long pumice-rich pyroclastic flows (F5; Fig. 2c). After the eruption, remobilization of the pyroclastic material generated lahars in main gullies around the crater.

The PDC deposits of the 1913 eruption of Colima were sampled at 7 sites (Fig. 1b) from the massive lithofacies (F4 and F5; Fig. 2c). Small clasts less than c. 2 cm in size were preferred because they reach thermal equilibrium more quickly and evenly than coarse clasts (Cioni et al. 2004), recording the actual temperature at deposition. A total of 25 larger clasts were also sampled and oriented in the field in order to derive the geographical orientation of the TRM component and compare the secondary direction to the Earth’s local magnetic field at the time of eruptions.

**Measurements and results**

Paterson et al. (2010) provided an exhaustive and up-to-date description of the methods, while Cioni et al. (2004) described the laboratory and interpretation procedures adopted in this study. Zanella et al. (in press) presented a concise account of the methods and discusses the use of oriented and non-oriented clasts, given that an accurate field orientation is not possible for fragments less than 2–3 cm in size.

A total of 192 specimens (86 at Colima, 106 at El Chichón) were measured at ALP Laboratory (Italy). The TRM components were studied using stepwise thermal demagnetization at 40 °C steps up to 580–600 °C. At each step, remanent magnetization was measured by a JR6 spinner magnetometer and the magnetic susceptibility checked by a kappabridge KLY-3 in order to detect possible chemical alteration due to heating. In order to improve the definition of the reheating temperature, fragments were split into 2–3 small bits whenever possible which were measured separately in four different runs shifted by 10 °C to each other.

The TRM components were analysed and isolated by principal component analysis (PCA; Kirschvink 1980) using the PaleoMac software (Cogné 2003). Only component directions defined with a maximum angular deviation (MAD) less than 10° were taken into account.

According to Cioni et al. (2004), the magnetic behaviour of individual clasts during thermal demagnetization can be divided into four main types (A, B, C and D) on the basis of the relation between the Tb spectra and the reheating temperature. These types reflect the differences in rock type, grain size of the magnetic minerals, thermal history and alteration from clast to clast. Type A and B clasts do not provide information because the whole of their Tb spectrum is above or below the reheating temperature, respectively. The temperature threshold between the two TRM components is well defined in type C clasts, whereas in type D it is somewhat hidden by secondary magnetization processes such as post-eruptive chemical alteration.

**Fig. 2.** Schematic stratigraphy (reconstructed) of the (a) 1982 eruption of El Chichón and (e) 1913 eruption of Colima volcanoes and position of the samples. Star = site from Sulpizio et al. (2008). Photographs: (b) PDC deposits of 1982 eruption of El Chichón at Esquipula Guayabal, site EG-07; (d) PDC deposits from Phases 2 and 3 along the San Antonio ravine and (e) PDC from Phases 1 and 2 along the Cordoban ravine.
\( M_{\text{max}} = 1.73 \text{ A m}^{-1} \)

\( M_{\text{max}} = 0.49 \text{ A m}^{-1} \)

Specimen: CHI9316-9

Specimen: CHI9316-7

(c)

Site: CHI9316

\[ T (°C) \quad 200° \quad 250° \quad 300° \quad 350° \quad 400° \quad 450° \quad 500° \]

| CHI9316-1 | CHI9316-2 | CHI9316-3 | CHI9316-4 | CHI9316-5 | CHI9316-6 | CHI9316-7 | CHI9316-8 | CHI9316-9 | CHI9316-10 |
|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|

\( T_{\text{dep}} \) 410–420 °C

primary TRM acquired during lava formation

low-T TRM acquired after 1982 PDC deposit formation
El Chichón

Most of the clasts sampled at El Chichón are of Cioni’s type C (Fig. 3). The reheating has little effect on the magnetization up to c. 400 °C. Thereafter, the intensity sharply decreases and the Zijderveld graphs point straight towards the origin. Most of the remanent magnetization is carried by the primary TRM component, stable above 440–450 °C, whereas a small secondary component is evident below. The curves are typical of type C clasts, which dominate among the El Chichón samples. The temperature threshold between the primary and secondary TRM is typically well defined within a range that corresponds to one heating step of 40 °C.

The reheating range of all clasts collected at a site was compiled and the site $T_{\text{dep}}$ assessed as the overlap of the maximum number of reheating ranges of individual clasts, as shown for site CHI9316 (Fig. 3c). $T_{\text{dep}}$ values of the El Chichón 1982 deposits vary from 360–370 °C to 410–420 °C with no substantial difference between the various depositional units and lithofacies (Table 1).

Colima

Thermal demagnetization diagrams of Colima samples show a somewhat more complicated thermal history (Fig. 4a, b). Clast VC4-5 shows the characteristic knee-shaped trend of type C; the curves are similar to those of El Chichón with the main difference being that the temperature threshold between the two TRM components is lower than those seen in Figure 3. The other clast, VC4-3, carries three TRM components with distinct $T_b$ spectra, which we interpret as evidence of two reheating processes. The high-temperature component ($T_b > 440$ °C) is the primary TRM acquired as the lava parent rock cooled. The clast was broken from the outcrop and eventually embedded within a PDC deposit of an eruption prior to 1913 and reheated to c. 440 °C. This overprint corresponds to the intermediate-temperature TRM component (440 °C > $T_b > 240–280$ °C; Fig. 4). Finally, a 1913 PDC reworked the clast and produced a second thermal overprint, recorded by the low-temperature component ($T_b < 240–280$ °C).

The reheating ranges of the clasts collected at site VC4 overlap at a $T_{\text{dep}}$ in the range 250–300 °C (Fig. 4c). The occurrence of two outliers with a higher reheating range of 340–380 °C as well as a reheating to 440 °C are consistent with the findings of Paterson et al. (2010). These authors interpret the very high dispersion of the low-temperature directions of the clasts they collected at Colima and the lack of any direction close to the Earth’s field at the time of the eruptions as the result of lahars remobilization of older pyroclastic debris, which implies that no thermal overprint occurred during deposition of the 1913 PDCs. On the other hand, many clasts have multiple TRM components which bear witness of a former reheating event at a temperature of the order 250–450 °C. The results of Paterson et al. (2010) and our results substantiate each other and show that the 1913 PDCs reworked pyroclastic debris laid down by older PDCs. The 1913 deposits at Colima have $T_{\text{dep}}$ values (Table 1) that vary between 250–280 °C and 310–330 °C.

The results of the magnetic measurements are summarized in Figure 5. The pie diagrams show that most clasts from the deposits of both volcanoes are type C of Cioni et al. (2004) and therefore yield reliable values of their reheating temperature. Some 15–25% are type D, whereas types A and B are rare and only occur at El Chichón. The $T_{\text{dep}}$ values do not

| Table 1. Deposition temperature of 1982 El Chichón and 1913 Colima PDC deposits |
|-----------------------------|-----------------------------|-----------------------------|
| Unit | Site | $n/N$ | $T_{\text{dep}}$ (°C) |
|-----------------------------|-----------------------------|-----------------------------|
| El Chichón                  |                             |                             |
| PS3                         | CHI9253                     | 9/12                        | 410–420                     |
|                            | CHI9256                     | 10/12                       | 410–420                     |
|                            | CHI9260                     | 10/11                       | 370–380                     |
|                            | CHI92289x                   | 11/12                       | 360–370                     |
|                            | CHI9314                     | 9/12                        | 380–390                     |
|                            | CHI9316                     | 10/12                       | 410–420                     |
|                            | BAF1                        | 10/12                       | 360–380                     |
|                            | CHI92123                    | 9/12                        | 410–420                     |
|                            | CH6-05*                     | 23/29                       | 360–400                     |
|                            | PS1                         | 21/24                       | 360–400                     |
|                            | CHI9285                     | 10/12                       | 360–380                     |
| Colima                      | Phase 3-F5                  |                             |                             |
|                            | VC4                          | 10/12                       | 250–280                     |
|                            | LB01                         | 10/12                       | 290–300                     |
|                            | CMS                          | 10/10                       | 280–300                     |
|                            | CMI                          | 9/9                         | 260–290                     |
|                            | Phase 2-F4                  |                             |                             |
|                            | CO13                         | 15/19                       | 280–300                     |
|                            | LB08                         | 11/12                       | 310–330                     |
|                            | LUM7                         | 12/12                       | 290–310                     |

$n/N$: number of successfully demagnetized specimens/number of collected specimens; *data from Sulpizio et al. (2008).
(a) $M/M_{\text{max}}$ vs. $T$ for Specimen VC4-5

$M_{\text{max}} = 0.32$ A m$^{-1}$

(b) Magnetic susceptibility vs. declination for Specimen VC4-3

$M_{\text{max}} = 0.71$ A m$^{-1}$

(c) Site: VC4

| Specimen | $T$ (°C) | 200° | 250° | 300° | 350° | 400° | 450° | 500° |
|----------|----------|------|------|------|------|------|------|------|
| VC4-1    |          |      |      |      |      |      |      |      |
| VC4-2    |          |      |      |      |      |      |      |      |
| VC4-3    |          |      |      |      |      |      |      |      |
| VC4-4    |          |      |      |      |      |      |      |      |
| VC4-5    |          |      |      |      |      |      |      |      |
| VC4-6    |          |      |      |      |      |      |      |      |
| VC4-7    |          |      |      |      |      |      |      |      |
| VC4-8    |          |      |      |      |      |      |      |      |
| VC4-9    |          |      |      |      |      |      |      |      |
| VC4-10   |          |      |      |      |      |      |      |      |
| VC4-11   |          |      |      |      |      |      |      |      |
| VC4-12   |          |      |      |      |      |      |      |      |

$T_{\text{vis}}$: 250–280 °C

Legend:
- primary TRM acquired during lava formation
- secondary intermediate-T TRM
- low-T TRM acquired after 1913 PDC deposit formation
show a definite relation to the depositional units and/or rock type at El Chichón, whereas at Colima the Phase 2 deposits have somewhat higher values that the Phase 3 deposits.

**TRM directions**

The model outlined in the previous sections assumes that the high-\(T\) TRM directions of the clasts are randomly distributed, whereas the low-\(T\) directions cluster around the Earth’s field direction at the time of the eruption. The analysis of the directions, however, comes up against two main difficulties:

1. field orientation of small clasts, which provide the more reliable values for the reheating temperature, is difficult or even impossible to determine; and
2. the definition of the low-\(T\) direction is commonly poor.

The first point is self-evident and the second is illustrated in the diagrams of Figure 4. The palaeomagnetic direction of a TRM component is given by interpolation of the corresponding points in the Zijderveld diagram. This interpolation becomes more accurate as the curve they outline grows longer, that is, as the intensity of the TRM component is larger. When the intensity is low, the points crowd together (Fig. 4a) and interpolation results in a poorly defined direction with a high maximum angular deviation value (MAD > 15°). On the other hand, when the intensity is high the points arrange in the curve over a greater length (Fig. 4b) and the interpolated direction is more significant (MAD < 10°).

A total of 15 clasts from Colima and 10 from El Chichón large enough to be accurately oriented in the field provided low-\(T\) directions with MAD < 10° (Fig. 6), which in our case can be considered as a reasonable cut-off value. The high-\(T\) directions are scattered as expected because the clasts lost their original orientation when broken out from the parent rock and chaotically transported within the pyroclastic cloud. On the other hand, the low-\(T\) directions cluster around the direction of the Earth’s field at the dates of the eruptions as derived from the International Geomagnetic Reference Field (IGRF) for 1913 and 1982, respectively (geomagnetic online calculator http://www.ngdc.noaa.gov/geomag/geomag.shtml). Moreover, their mean directions are statistically indistinguishable from the IGRF directions. The Colima mean direction is \(D = 4.5°, I = 43.1°\) (Fisher’s statistics parameters \(k = 38, \alpha_{95} = 6.3°\)) compared to the IGRF values of \(D = 9.3°, I = 44.7°\). For El Chichón the mean direction is \(D = 352.1°, I = 40.2°\) (\(k = 39, \alpha_{95} = 11.2°\)) compared to the IGRF values of \(D = 5.8°, I = 45.8°\). The good agreement between the directions of the low-\(T\) TRM component and the IGRF is robust evidence that the TRM was acquired when the deposit cooled.

**Discussion**

**Comparison with previous data**

The areal distribution of the estimated \(T_{dep}\) values of the El Chichón 1982 PDC deposits studied in this paper, together with those of Sulpizio et al. Fig. 4. Thermal demagnetization diagrams for Colima samples. (a) Normalized intensity decay curve and (b) Zijderveld diagrams. Dark grey: primary high-\(T\) TRM; middle grey: secondary intermediate-\(T\) TRM; light grey: secondary low-\(T\) TRM; others symbols as in Figure 3. (c) Estimate of the site deposition temperature \(T_{dep}\) by overlapping of reheating ranges of individual clasts.
are consistent with the maximum temperature of 402 °C measured a few days after the eruption by a thermocouple inserted into the deposit 5 km NE of the vent (Sulpizio et al. 2008; Fig. 7). The $T_{\text{dep}}$ values depend on neither the lithological units and lithofacies (Fig. 2, Table 1) nor the geographic location with respect to vent. We can therefore conclude that the $T_{\text{dep}}$ values are similar throughout the eruptive succession, with no significant variation either in space or in time.

$T_{\text{dep}}$ was previously estimated at Colima volcano for a debris avalanche (Clement et al. 1993) and for PDC deposits of 1913 eruption (Paterson et al. 2010). High $T_{\text{dep}}$ values were reported for the debris avalanche deposits (c. 350 °C), although a strong variability was observed with respect to distance travelled and other physical parameters (e.g. the deposit thickness). The high $T_{\text{dep}}$ was interpreted as suggestive of a simultaneous occurrence of the debris avalanche and a magmatic eruption (Clement et al. 1993). Paterson et al. (2010) reported very low $T_{\text{dep}}$ and random orientation of the magnetic components from samples of the 1913, 2004 and 2005 PDC deposits. These data were interpreted as due to reworking of primary PDC deposits from volcaniclastic flows (i.e.

\[
\begin{align*}
\theta &= 15.0 \pm 4.5^\circ, \phi = 43.1^\circ, \eta = 38.6^\circ, 6.3^\circ \\
\text{IGRF 1982} &
\end{align*}
\]

\[
\begin{align*}
\theta &= 10.0 \pm 52.1^\circ, \phi = 40.2^\circ, \eta = 39.1^\circ, 11.2^\circ \\
\text{IGRF 1982} &
\end{align*}
\]

Fig. 6. Equal area projection of TRM directions from oriented specimens from Colima (left) and El Chichón (right). Dot: low-$T$ direction; star: mean low-$T$ direction with 95% confidence ellipse; diamond: IRGF direction at the date of eruption.

Fig. 7. Areal distribution of deposition temperature $T_{\text{dep}}$ at El Chichón volcano. The location of the maximum $T$ measured 4 days after the eruption using a thermocouple is indicated in yellow.
lahars), and therefore not representing the $T_{\text{dep}}$ of the original PDCs from 1913, 2004 and 2005 eruptions (Paterson et al. 2010). The new $T_{\text{dep}}$ reported here (Fig. 8) highlight magmatic temperatures for all the analysed samples, confirming the probable secondary deposition of the samples from 1913 eruption reported by Paterson et al. (2010).

Comparison of $T_{\text{dep}}$ of El Chichón and Colima PDCs

The investigation of $T_{\text{dep}}$ in PDC deposits from the 1982 eruption of El Chichón and 1913 eruption of Colima volcanoes highlight two different ranges of $T_{\text{dep}}$, despite similar magma composition and lithofacies of PDC deposits (Fig. 9). The two eruptions also show similar dispersal of data (black bars in Fig. 9c, f) and half-dispersal data (defined as the $T_{\text{dep}}$ range assessed using the smaller $T_{\text{dep}}$ ranges from at least 50% of analysed specimens from each sample; red bars in Figure 9c, f; supplementary material). This indicates similar behaviour of re-heating of analysed lithic material and the TRM data.

It has been argued that transport mechanisms and sedimentation/deposition processes can influence the emplacement temperature of PDCs (Cioni et al. 2004; Sulpizio et al. 2008; Zanella et al. in press). Although this in principle is correct, the effect of sedimentary processes on emplacement temperature is appreciable only when fluid turbulence is greatly enhanced by surface roughness (Gurioli et al. 2005; Zanella et al. 2007, in press). Lithofacies analysis of 1982 El Chichón (Sulpizio et al. 2008) and 1913 Colima (Saucedo et al. 2010) PDC deposits suggest they were emplaced under similar flow-boundary conditions, in which turbulence was inhibited by high particle concentration in the lower flow boundary. This rules out any significant difference in transport and deposition mechanism in determining the reason for the difference in $T_{\text{dep}}$ between the two eruptions.

Possible explanations have to be investigated in the light of eruptive $T$ and/or pre-eruptive fugacity of oxygen that influenced the formation and composition of magnetic oxides of lava fragments used for determining $T_{\text{dep}}$.

Prior to the 1982 eruption of El Chichón, the magma (of trachyandesitic composition) equilibrated at a pressure of c. 2 kb (Luhr 1990) and at an estimated temperature of 750–850 °C (Luhr et al. 1984; Rye et al. 1984; Luhr 1990). These estimates agree with the pre-eruptive temperature.
of 820–830 °C (oxygen fugacities of −11.08 to −11.02 and H₂O contents of 5–6 wt%) for the 550 year BP El Chichón eruption, whose magma equilibrated at pressures of c. 2–2.5 kb (c. 6–7.5 km) with the crystallization of plagioclase, hornblende, clinopyroxene, magnetite and rare ilmenite (Macías et al. 2003). The magma of the 1913 eruption of Colima had an andesitic composition, with a mineralogical assemblage consisting of plagioclase > orthopyroxene > clinopyroxene > hornblende + magnetite-ilmenite + apatite and rare resorbed olivine. Pre-eruptive temperature was below 950 °C, based upon the presence of amphibole (Savov et al. 2008). Laboratory experiments suggest that amphibole is stable at pressures c. 200 Mpa (c. 6 km depth) and at temperatures of c. 860 °C (Macías et al. 2013). These mineralogical and petrological data indicate how pre-eruptive conditions cannot explain the differences in $T_{\text{dep}}$ of 1982 El Chichón and 1913 Colima PDC deposits.
Conclusions

The measurement of $T_{dep}$ of PDC deposits from the 1982 eruption of El Chichón and the 1913 eruption of Colima volcanoes provided new data on equilibrium temperature of pyroclastic deposits from these two important Mexican volcanoes. These data are important for better understanding the physics of PDCs and for improvement of hazard mitigation strategies. In particular, measured $T_{dep}$ show the limited influence of flow-boundary conditions at the time of deposition in determining the equilibrium temperature of lithic clasts with deposit. This is because similar $T_{dep}$ have been measured within each eruption for different lithofacies and at different distances from the vent.

Similar magma composition and pre-eruptive conditions rule out any influence of magmatic parameters in the observed difference in measured $T_{dep}$ between the El Chichón and Colima PDC deposits. These results highlight our partial understanding of mechanisms that influence the attainment of thermal equilibrium between the heat carriers (juvenile particles and gas) and embedded (cold) lithic fragments. Further and more detailed laboratory and field research is required.

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