Hydrovolcanic eruptions are common in coastal environments where erupting magma may interact with seawater in either subaerial or subaqueous settings on the location of the eruption conduit and vent (Sheridan & Wohletz 1983). Nearshore and shallow subaqueous activity may generate high-intensity explosions characterized by ballistic blocks, ash-fall and pyroclastic density currents, which, in contrast to the lesser impact of magmatic eruptions of the same magnitude, may severely affect surrounding areas. This type of activity is an example of the major role played by the hydrogeological setting in determining the eruption style throughout the entire eruption. For example, phreatomagmatic volcanoes are common in basin or coastal areas as well as in fault zones and other fracture systems that host groundwater (Lesti et al. 2008), whereas spatter cones and lava flows occur in topographically higher locations (White 1991; Sohn & Park 2005). One of the main issues concerning hydrovolcanism is that although the eruption of a single monogenetic landform might be fed by small magma volumes, the host environment exerts a strong influence on the style of eruption (Lorenz 2003; Ross et al. 2011).

The greatest hazard to life on basaltic volcanic islands (such as Deception Island) is from explosive phreatomagmatic eruptions (Németh & Cronin 2009) and this type of volcanism, occurring on a volcanic island with changing subaerial conditions (i.e. fracture system and availability of external water) has been documented in several places worldwide: Taveuni volcano, Fiji (Cronin & Neall 2000, 2001; Cronin et al. 2001); West Ambrym, Vanuatu (Németh & Cronin 2007); Ambae Island, Vanuatu (Németh & Cronin 2009); Ambrym Volcano, Vanuatu (Németh & Cronin 2011); Miyakejima Volcano, Japan (Geshi et al. 2011); Montañita Pelada Tuff Ring, Tenerife (Carmona et al. 2011); El Golfo Tuff Cone, Lanzarote (Pedrazzi et al. 2013).

In this paper we report a study of the 1970 eruption at Deception Island, the last volcanic episode of the South Shetland Archipelago, outlining how the near-surface environment parameters (groundwater availability, fracture system and country rock properties) have a great potential to influence eruption processes and related hazards.

The South Shetland Islands volcanic arc (Antarctica) is the result of the subduction of the Phoenix plate beneath the Antarctic plate at the South Shetland trench (Saunders et al. 1982; Smellie et al. 1984; Smellie 1990). The subsequent opening of the Bransfield Strait is probably related to passive subduction of the former Phoenix plate and roll-back at the South Shetland trench (Barker 1982; Lawver et al. 1995). The basin is volcanically active and several subaerial and submarine volcanic edifices are recognized along the major rifting axis. The three main volcanic Quaternary edifices of the archipelago are Deception, Bridgean and Penguín Islands, and numerous small volcanic centres are recognized also on Livingston, Greenwich and King George Islands (Weaver et al. 1979; Smellie et al. 1984, 1995; Fisk 1990; Keller & Fisk 1992; Gracia et al. 1996).

The 1970 eruption is an excellent example of monogenetic eruption in a coastal environment (e.g. Sheridan & Wohletz 1983; Sohn et al. 2008, 2012; Carmona et al. 2011). We refer to the 1970 eruption as one eruptive episode or eruptive event with different eruptive phases. This eruption took place in the northern part of the Deception caldera and gave rise to the formation of at least 13 vents that differ significantly in the shape of their craters (Baker & McReath 1971, 1975). Several preliminary accounts of the 1970 eruption have already been published (e.g. Baker & McReath 1971, 1975; González-Ferrán et al. 1971; Shultz 1972). However, a detailed description of the succession of deposits, as well as an accurate reconstruction of the eruptive activity, is still lacking.

In this paper we describe in detail the stratigraphy, lithology, and granulometric and depositional characteristics of the deposits of the 1970 eruption and use this information to infer the eruption dynamics.
and the transport and depositional processes, noting how this type of activity is strongly influenced by the shallow host conditions such as groundwater availability, country rock properties, and fracture and fault system. We then compare these data with those from the 1967 eruption, analysing the hazard implications of this type of eruptive activity on Deception Island, which hosts temporary military bases and scientific installations and is becoming one of the main tourist attractions in Antarctica, with almost 50000 people visiting the island in 2004–2010 (IAATO, International Association of Antarctica Tour Operators, http://iaato.org/tourism-statistics).

Geological setting and general characteristics of Deception Island

Deception Island is the southernmost island in the South Shetland Archipelago and is located at the southwestern end of the Bransfield Strait, a young (<1.4 Ma) back-arc basin (Fig. 1a), which was formed as a consequence of the Phoenix plate subduction under the Antarctic plate during the late Mesozoic–Cenozoic interval (Dalziel 1984). The Bransfield basin, <60 km wide and 500 km long, separates the South Shetland Islands from the Antarctic Peninsula (Fig. 1b) and has a characteristic graben structure, with tilted blocks and rotational faults developed under a regime of continental extension (Jeffers & Anderson 1990; Gracia et al. 1996; Vuan et al. 2005). Deception Island consists of a ~0.75 Ma (Valencio et al. 1979; Smellie 1988) horseshoe-shaped volcanic edifice, interpreted as a tectonic fault-controlled collapse caldera (Marti et al. 2013), with a below-sea-level diameter, determined through multibeam bathymetry, of 25 km and an above-sea-level diameter of around 13 km (Smellie 1988; Barclay et al. 2009; Fig. 1c). A sea-flooded depression known as Port Foster occupies the central part of the island (Smellie 2001; Marti et al. 2013).

Deception Island is the most active volcano in the South Shetland Archipelago (Antarctic Peninsula group) and over 20 eruptions have been identified over the past two centuries (Roobol 1980; Pallas et al. 2001; Smellie et al. 2002). The oldest episodes were inferred from topographical changes and historical accounts (Pallas et al. 2001) suggesting different types of activity with hydromagmatic (e.g. Crater Lake and associated tuff cones, pre-1829; Kroner Lake, 1830–1909) and magmatic eruptions (cinder cones north of Mount Kirkwood, 1839–1842) (Fig. 1c). The most recent episodes occurred in 1967, 1969 and 1970 (see Orheim 1971; Baker et al. 1975; Roobol 1980; Smellie 2002; Fig. 1c).

The construction of this composite volcano, which rises to 540 m above sea level at Mount Pond and 460 m at Mount Kirkwood (Fig. 1c), was truncated by the formation of a central caldera (Port Foster) resulting from an explosive eruption of basaltic to andesitic magmas, mostly as pyroclastic density currents with a total pyroclastic bulk volume of the order of 90 km³ (Marti et al. 2013). The caldera has a polygonal shape and was controlled by faults related to the regional tectonics (Marti et al. 2013). The basement of the island is not exposed but probably corresponds to Cretaceous–Tertiary sedimentary and volcanic rocks and/or unconsolidated and poorly consolidated pre-Quaternary marine sediments deposited on the submerged South Shetland Islands platform (Ashcroft 1972; Grad et al. 1992; Smellie 2001). The pre-caldera units (Basalt Shield Formation; Fig. 2) consist mostly of various rock units derived from a shield volcano, in which different eruptive cycles and vents can be distinguished (Smellie 2001). Most of these units correspond to lava flows and Strombolian deposits, as well as to palagonitized hyalo-clast breccia. The shield-related units are overlain unconformably by a thick sequence of massive ignimbrites and minor pyroclastic surge deposits of basaltic to andesitic composition, which represent the syn-caldera deposits (Marti et al. 2013) (Outer Coast Tuff Formation; Fig. 2). This succession is overlain unconformably by discontinuous sequences of post-caldera deposits originating from different vents (Post-Caldera Deposits; Fig. 2), most of them located inside or around the caldera depression itself (Smellie 2001).

Methods

A photointerpretation of orthophotos at a scale of 1:5000 was necessary to identify the potentially most interesting outcrops in the area. Fieldwork was conducted during the austral summer of 2012–2013 in an area of about 5 km² around the cones of the 1970 and 1967 eruptions. A total of six detailed and representative stratigraphic logs were established but, owing to the difficult access to some zones (mainly because of weather conditions and ice cover), it was not possible to cover the whole area that preserves the deposits of these eruptions. Nevertheless, the stratigraphic logs provided are representative of the resulting successions of deposits. Log 1 represents most probably an overlapping sequence of deposits from multiple single vents in the eastern part of the area, allowing us to see the overall stratigraphy, and Logs 2–6 illustrate in detail the eruptive sequence of the most representative craters of the western part.

The geometry and lithology of the deposits were documented at each outcrop following lithostratigraphic and stratigraphic criteria; the colour, nature and relative content of components, grain-size variations, texture and depositional structures were analysed. Nomenclature used in the text for bed thickness, grain size, and sorting of the pyroclastic deposits follows that proposed by Sohn & Chough (1989). Maximum class sizes (scoria and lithic fragments) of each identified layer were established based on the geometric mean average of the three axes of the five largest clasts, from hand samples or in some cases from scaled photographs of the outcrops from a horizontal area of 0.5 m².

All data were managed using Surfer 7 and an open-source geographic information system (GIS) framework (QGIS, www.qgis.org). The location of study sites and the position of the stratigraphic logs were recorded using a portable Garmin Dakota 20 global positioning system (GPS) with a precision of about 3 m, using Global Mapper professional software to manage waypoints. The reference zone used was the UTM projection Datum: D_WGS_1984, zone 20 South.

Grain-size analyses employing dry-sieving techniques and component analysis were performed by weighing or counting 42 representative samples. Samples were sieved at 1σ intervals using sieves with mesh sizes in the range 32–1/32 mm (−5φ to 5φ). Clast compositions were initially characterized in the field via hand-sample observations and then later in the laboratory under a binocular microscope. Component analysis was carried out on the −5φ, −4φ, −3φ, −2φ and −1φ fractions of the deposits. Clasts were separated into three classes: juvenile (scoria) and two types of accidental lithic fragments (ignimbrite and lava clasts). Further analysis of the samples larger than −1φ is not provided here because they were not considered statistically significant.

The physical parameters of the eruption were obtained: the erupted volume was calculated with an isopach map obtained from Baker & McReath (1975) and Smellie (2002) using the method of Bonadonna & Costa (2013), and the plume height was determined by applying the same method to a median grain-size diameter (Mdφ) isoline map from the data of Geyer et al. (2006, 2008).

Characteristics of the 1970 craters

The 1970 craters are located in the northern part of Deception Island, between Goddard Hill and Cross Hill, at the foot of the caldera wall and close to the beach (Fig. 3a). At least 13 vents from this eruption were identified and can be separated into two main groups, the eastern and western, based on the location and shape of the craters. The eastern group is located at the foot of Goddard Hill and consists of seven
conical edifices (a–g; Fig. 3b) with a maximum depth of 150 m, aligned roughly NW–SE (Fig. 3b), Crater g, described by Baker & McReath (1975), has been eroded away by the island’s glacier. The western group, with ‘maar-like’ shapes (i.e. deep craters with steep crater walls that cut into the syn-eruptive surface (Ollier 1967; Lorenz 1973; Fisher & Schmincke 1984)), is located at the foot of the exposed part of the caldera wall (h–m; Fig. 3b); its craters have almost vertical sides with cliffs that reach 20–25 m in height. The first crater (h, i; Fig. 3b) is a composite structure consisting of two major craters whose southern-most points are connected to Telefon Bay by a narrow isthmus c. 10 m wide. The second crater (j; Fig. 3b), a depression located in the north-western part of the group originating from the 1967 eruption, is partially filled by a lake and seems to represent the remains of another eruptive centre. The last three craters (k–m; Fig. 3b) are connected to each other

Fig. 1. (a) Simplified regional tectonic map and location of the South Shetland Islands Archipelago (modified from Ibáñez et al. 2003). HFZ, Hero Fracture Zone; SFZ, Shackleton Fracture Zone. (b) South Shetland Islands Archipelago and location of Deception Island (modified from Grad et al. 1992). (c) Orthophotomap of Deception Island (http://lagc.uca.es/web_lagc/orfo.jpg). Sites of 1967, 1969 and 1970 events are also shown.
by narrow channels and to Port Foster by a 150 m wide channel located in the eastern part.

**Characteristics of the pyroclastic succession**

We carried out a detailed characterization of six stratigraphic logs to reconstruct the succession of deposits. Although limited exposure allowed us to obtain only one representative stratigraphic log from the Eastern Craters (1; Fig. 4), five stratigraphic logs were obtained from the Western Craters (Lago Escondido Edifice) (2–6; Fig. 4).

**Eastern craters**

Stratigraphic log 1 was described between crater b and c (Fig. 3b) and represents a total thickness of about 40 m with an irregular angular unconformity in the middle (Figs 4 and 5a, b), marking a visible change between the two parts of the succession (22–41 m). The base of the succession is not exposed (Figs 4 and 5e). The first part of the stratigraphic log (0–1 m; Fig. 4) shows a continuous change of laminated poorly sorted medium-thick (10–30 cm) beds, from fine to coarse vesiculated lapilli-sized clasts with accidental lithic fragments (syn-caldera ignimbrites and lava flows). This deposit suddenly changes to poorly sorted, very thick (>100 cm) lithic-rich breccia beds with poorly vesiculated bombs (1–4 m; Fig. 4). The succession continues from 4 to 11 m with a sequence of very thick (>100 cm) bomb-rich beds (Figs 4 and 5c), alternating with massive, poorly sorted, lithic-rich very thick (>100 cm) breccia beds (Figs 4 and 5d). The next deposit is about 50 cm thick, has an erosive base, and is made up of coarse lapilli and bombs with a yellow alteration (Fig. 5a), which implies that at the moment of deposition it might have retained a greater proportion of water as steam compared with the other deposits. The rest of the first half of the succession (11–22 m; Fig. 4) consists of poorly sorted, very thick (>100 cm) lithic-rich (lava and ignimbrite clasts) breccia deposits without any visible depositional structures (Fig. 4). A marked unconformity indicates a change in the type of deposits in

Fig. 2. (a) Simplified geological map of Deception Island (modified from Martí *et al.* 2013) showing the isopachs (dashed lines; in centimetres) of the 1970 eruption (modified from Baker & McReath 1975); (b) isopachs (dashed lines; in centimetres) of the 1970 eruption in the South Shetland Archipelago (modified from Baker & McReath 1975).
the second half of the succession (Figs 4 and 5a, b), which is characterized by a continuous alternation of thick (30–100 cm) to very thick (>100 cm), coarse well-sorted lapilli-sized breccia beds with accidental lithic fragments and non-vesiculated (juvenile) scoria and medium-thick (10–30 cm) fine and coarse lapilli beds with planar stratification (Fig. 4).

**Western Craters**

Stratigraphic logs 2–6 were described around the crater of Lago Escondido, the easternmost edifice of the Western Craters (h, i; Figs 3b and 5f). All stratigraphic logs show a similar succession, mainly characterized by very thick (>100 cm), poorly sorted, lithic-rich (ignimbrites and lava clasts) breccia beds with some medium-thick (10–30 cm) beds of poorly sorted coarse lapilli with weak lamination (Figs 4 and 5g, h). The thickness of the sequence varies from about 6 m in stratigraphic log 2 to less than 15 m in stratigraphic log 6 (Fig. 4). A crude stratification can be identified in some of the deposits (stratigraphic logs 2 and 6; Fig. 4), as well as enrichment of scoria bombs (stratigraphic log 2; Figs 4 and 5i). A general increase in the content of lithic clasts and grain size is observed in the uppermost deposits (Figs 4 and 5j).

**Distribution of the finest deposits**

All vents were located in the northern part of the island and the ejecta were mostly carried northwards. Ash deposits were found to the northern sector of the island between the central part of Kendall Terrace and Pendulum Cove (Fig. 2a). No evidence of ash was found on Mount Kirkwood. A considerable amount of the ash fell beyond Deception Island, as shown in Figure 2b, and c. 4 mm of fine ash fell on Arturo Prat station, Greenwich Island, and about 1 mm on Bellingshausen station and King George Island (Baker & McReath 1975). As reported by Baker & McReath (1971, 1975) and shown in Figure 2b, variations in thickness and distribution are consistent with a northeastward pattern of isopachs that extends along the axis.
Fig. 4. Composite stratigraphic logs of the deposits (log numbers refer to the position in Fig. 3b) and component analysis of the most representative samples. Stratigraphic log 1 refers to the Eastern Craters, and stratigraphic logs 2–6 are located around Lago Escondido. Black numbers above the columns refer to UTM coordinates (WGS 1984 Zone 20 S). Vertical variations in grain size and the maximum diameter of lithic and scoria clasts are also indicated.

of the South Shetland Islands under the influence of a southwesterly wind, which carried part of the material into the Bransfield Strait.

Grain-size, modal variation and clast distribution

Vertical variations in the grain-size distribution and modal variations were analysed by selecting representative samples from both the coarse- and fine-grained layers (Fig. 4). The vertical variation of maximum clast sizes (bombs and blocks) is shown in Figure 4. At stratigraphic log 1 there is a dominance of coarse deposits in the lower part with a general decrease in grain size throughout the succession. The median diameter (Mdφ) values of all samples range from −2.9 to −0.15, and sorting (σφ) values range from 1.08 to 3.76 (Fig. 6a). Deposits from the lowest part of the succession are poorly sorted compared with those from the upper part, which are generally finer grained. Skewness (αφ) varies from 0.27 to −0.75 (Fig. 6b). In the general trend of the lithic clast distribution (Fig. 4), the lowest part of the succession has the largest proportion of blocks, which can measure up to 60 cm.

Stratigraphic logs 2–6 have similar successions. A dominance of coarse deposits is observable in all the stratigraphic logs, with just an occasional slight decrease in grain size corresponding to coarse and fine lapilli layers. The median diameter (Mdφ) values of all the samples range from −3.11 to −0.018, and sorting (σφ) values range from 1.75 to 3.52 (Fig. 6a); the deposits are generally poorly sorted. Skewness (αφ) varies from 0.34 to −1.96 (Fig. 6b). Maximum clast size reaches 60 cm except in the upper part of the succession, where a general increase of up to several decimetres is observed (Figs 4 and 5j).
Figure 6c and d illustrates the grain-size distribution of the samples shown in Figure 4 and highlights the fact that the most representative classes range from $0\phi$ to $-3\phi$. An isopleth map from close to the eruptive centres is given in Figure 7. The size values recorded on this map represent an average of the three axes of the largest clasts from any given site. Clasts are made of juvenile fragments and accidental lithic fragments (grey aphyric lava, yellow and greenish tuff and reddish vesicular lava). Some of the lithic clasts related to the Western Craters that are shown in Figure 7d might have been affected by redistribution by streams and mudflows resulting from the unusual seasonal melt on Deception Island during the 1967–1968 melt, as described by Baker & McReath (1975). An isolated area of relatively large bombs between Cross Hill and Wensleydale Beacon with no traces of a source crater was reported by Baker & McReath (1975). These bombs were probably discharged from another eruptive vent, as shown by the isopleth map of Figure 7a.

A preliminary median grain-size diameter ($M_{d0}$) isoline map for the distal deposits (Fig. 7e) was obtained from data reported by Pallás et al. (2001), Fretzdorff & Smellie (2002) and Geyer et al. (2006, 2008). Further data were taken from Hodgson et al. (1998), who reported an ash layer in a core from Midge Lake (Livingston Island). Hodgson et al. (1998) suggested that this horizon might correspond to any of the documented eruptions on Deception Island (1967, 1969 or 1970) or to the earlier eruptions that occurred in 1842 and 1912–1917. However, the reported dispersal area of the 1967 and 1969 eruptions (Baker & McReath 1975; Smellie 2002) and the shallow position in the core of this ash layer suggest that this horizon belongs to the 1970 eruption, although the 1842 and 1912–1917 episodes, which were only mentioned briefly by Wilkes (1845), Orheim (1971) and Roobol (1973, 1980, 1982), cannot be ruled out as possible origins.

**Component analysis**

The products of the 1970 eruption consist of a mixture of black dense or vesicular scoria of basaltic andesitic composition with differing proportions of accidental lithic clasts (older lava flows and yellow and greenish tuff) (Fig. 4). Variations in the occurrence of the lithic fragments can be seen in the stratigraphic succession, as well as in the two groups of craters. As observed in Figure 4, the lava clasts are the most abundant accessory components in the successions in both groups of craters. In the Lago Escondido edifice, the lava clasts are particularly dominant in the lower part of the succession (e.g. samples 4-A4, 4-A5, 4-B1, 4-C1, 4-E2, 4-E6; Fig. 4) and also characterize the whole succession in stratigraphic log 1 (Fig. 4). Greenish and yellow tuff clasts are typically present in smaller number and in general are found in the lower part of stratigraphic log 1 (e.g. samples 3-A1; Fig. 4) and in the upper part of the sequence from Lago Escondido (e.g. samples 4-E7, 4-F5; Fig. 4).

Generally, accidental lithic clasts increase in percentage in the size range $-4\phi$ to $-3\phi$ (Fig. 4). Small variations in the amount of juvenile components can be observed throughout the first outcrop (eastern group), where the juvenile content ranges around 90%. The second group (western group) of craters shows a general decrease in the amount of juvenile components of up to 70–80% (Fig. 4).

**Discussion**

Despite the limited number of exposures that we were able to access, we managed to establish one stratigraphic log from the Eastern Craters (stratigraphic log 1; Fig. 4) and five from the Western Craters from around the same vent area (Lago Escondido Edifice) (stratigraphic logs 2–6; Fig. 4). We believe they are representative of the complete succession of deposits formed by the 1970 eruption and its dynamics.

This eruption seems to have been continuous, although the presence of eruptive dynamics (as hereinafter explained), and was totally pyroclastic (magmatic and phreatomagmatic) with no lava flow. The location of the craters along one of the main structural limits of the caldera (see Martí et al. 2013) suggests that caldera faults were probably guiding the rising magma to the surface. Several vents opened during the eruption and generated different types of craters and cones with contrasting eruptive styles.

The Eastern Craters near Goddard Hill contrast with the Western Group at Telefon Bay, which consists of maar-like craters. As suggested by Sohn (1996), the morphology of edifices might be controlled by depositional processes; nevertheless, the shape of the landform will also depend on the physical properties of the surrounding bedrock (pre-eruptive surface), vent geometry and the
Fig. 5. Field photographs of the main stratigraphic logs. (a) General view of outcrop 1 showing the unconformity (dashed white line) in its highest part. Dotted and dashed black lines indicate, respectively, the dip of the layers (yellow altered deposit) below and above the unconformity. Black circle indicates the person used as a scale. (b) Detailed view of outcrop 1. The lower part (black dotted lines) is characterized mainly by massive deposits, whereas the upper part above the unconformity (dashed white line) is formed by alternating (black dashed lines) thick, coarse, lapilli-sized breccia layers and fine and coarse lapilli layers with planar stratification. (c) Example of bomb-enriched deposits. (d) Typical massive lithic-rich deposit; the scraper is about 15 cm long. (e) General view of the inner part of the easternmost crater, with a height of about 20 m in the background. (f) General view of Lago Escondido. (g) Typical outcrop at Lago Escondido, consisting mainly of massive, unbedded and subordinate fine lapilli deposits; the shovel is about 1 m long. (h) Detail of a massive deposits; the stick is 10 cm long. (i) Bomb-rich deposit. (j) Example of lithic clast close to Lago Escondido Craters. L, lava; I, ignimbrite; S, scoria.
The water/magma ratio (Lorenz 2003; Auer et al. 2007; Basile & Chauvet 2009; Martin-Serrano et al. 2009; Ross et al. 2011; White & Ross 2011; Sottili et al. 2012).

The lithological and depositional characteristics of the deposits forming the stratigraphic succession in the eastern cones reveal that the lowest part corresponds to deposits formed from base-surge-type explosions (e.g. Druitt 1998; Branney & Kokelaar 2002; Dellino et al. 2004a,b; Brand & White 2007; Brand & Clarke 2009). The high degree of fragmentation of these deposits could be an indication of phreatomagmatic activity with optimal magma–water energy transfer. Furthermore, fine-grained deposits might be the result of magma–water interaction with a soft unconsolidated substrate, which in this case might be represented by the shallow altered deposits of the Outer Coast Tuff Formation or by the heterogeneities in the same syn-caldera deposits, which are characterized by hard consolidated ignimbrite but also by well-stratified, lapilli to ash deposits (Martí et al. 2013). Although the deposits may appear to be formed by high-energy fragmentation, they might be the result of a low-energy outburst that displaced the unconsolidated substrate. Furthermore, as proposed by Tuffen et al. (2002) for the rhyolitic tuya at SE Rauðufossafjöll, Torfajökull, Iceland, a first sub-glacial phase of the eruption might have occurred, with explosive magma–water interaction generating fine-grained deposits within a well-drained ice vault followed by breaching of the glacier surface and involving the shield and the syn-caldera deposits (Fig. 8a1), as indicated by the presence of both types of rocks as accidental lithic clasts in these breccia deposits, which mostly correspond to shield-derived lava fragments and syn-caldera ignimbrites, suggests that at this stage the locus of the eruption was located in a part of the shield structure that was below the syn-caldera deposits (Fig. 8a2). In fact, as the eruption progressed, the hydromagmatic blasts could have disrupted a possibly fractured aquifer in the shield formation, enhancing a secondary permeability and thus causing an increase in the hydraulic conductivity as the water/magma ratio in the system increased dramatically, reaching the condition for the formation and emplacement of the breccia deposits. This stage of the eruption would be similar to those described by Aranda-Gómez & Luhr (1996) at Joya Honda maar, Mexico, by Németh et al. (2001) at Tihany Maar, Hungary, or by Pedrazzi et al. (2014) at Sant Dalmai, Spain.

A significant unconformity (Figs 4 and 5a, b) marks the transition towards the upper part of the succession, which is dominated by alternating thick, coarse, well-sorted lapilli-sized breccia layers and fine and coarse lapilli layers with planar stratification, indicating the pulsatory nature of the eruptions as well as the varied and contrasting rheological architecture of the country rocks. The presence of this unconformity, marked by a change in the layer dip, might suggest a shift in the locus of the explosions or vent widening, which is common at basaltic phreatomagmatic volcanoes (Sohn & Park 2005): the lower part of the succession could belong to crater c and the upper part to crater d (Fig. 2). As suggested by Sohn et al. (2012), these processes may also relate to changes in the magma supply.

Unlike the craters near Goddard Hill, the activity in the western part started with a submarine eruption (Fig. 8b1), as deduced by the actual position of the coastline compared with the 1969 coastline.

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**Fig. 6.** Plots of grain-size data from the fall and surge deposits. (a) Sorting (σφ) v. median diameter (Mdφ); (b) sorting (σφ) v. skewness (αφ); (c, d) granulometrical frequency distribution at (c) Eastern Craters and (d) Western Craters.
(see Shultz 1972; Baker & McReath 1975), and then evolved into a subaerial phase (Fig. 8b2, b3), leading to the formation of a new strip of land about 1700 m long, 400 m wide and 12 m above sea level. The 1970 activity caused the partial destruction of the island that was formed in 1967 in Telefon Bay, which was truncated at its northeastern and southwestern ends (Baker & McReath 1975). The stratigraphic logs (2–6; Fig. 4) studied in this sector were all located around the same vent area (Lago Escondido), whose morphology indicates the presence of two partially coalesced craters. The succession of deposits observed in all of these craters is very similar and is mainly characterized by breccia deposits. Stratigraphic logs 2 and 4 (Fig. 4) reveal the presence at the base of the breccia deposits of a lithic-poor layer (sample 4-A1; Fig. 4), which might correspond to the first submarine phase of the eruption (Fig. 8b1). This episode seems somewhat similar to those of Ambrin Volcano, where on December 1913 a phreatomagmatic eruption occurred along a rift system and the crater was initially excavated over 16 m below sea level and formed a lagoon (Németh & Cronin 2011), and Ambae Island, where phreatomagmatic activity on 28 November 2005 started with a submarine eruption where the rift axis meets the sea (Németh & Cronin 2011). Stratigraphic log 6 (Fig. 4) has a uniform succession of breccia deposits and subordinate fine lapilli deposits emplaced by pyroclastic surges, which increase in number in the upper part of the sequence (Fig. 4). The lithic content suggests an initial interaction with lava flows, probably emplaced from the post-caldera phase (e.g. samples 4-A4, 4-A5, 4-B1, 4-C1, 4-E2, 4-E6; Fig. 4) (Fig. 8b2), with a change in the upper part of the sequence characterized by an increase in the number of yellow tuff lithic clasts (e.g. samples 4-E7, 4-F5; Fig. 4) in both craters. Following the model proposed by Lorenz (1986), in the down-migrating explosion locus and gradual deeper excavation model, a progressive lowering of the position of the fragmentation level in the eruption conduit during the course of the eruption might have led the magma to interact with a deeper part of the basement made of syn-caldera yellow tuff (Fig. 8b3), similarly to normal maar volcanism as proposed by White & Ross (2011).
As reported by Baker & McReath (1975) this eruption occurred during the austral winter, when Deception Island was unoccupied. Preliminary information about the eruption came from the Chilean station Arturo Prat on Greenwich Island, the Soviet station Bellinghausen on King George Island and also from the British Antarctic Survey on 12 and 13 August 1970, but there were no eyewitnesses to the event.

In any case, on the basis of the available field data, a preliminary quantitative evaluation of the eruptive parameters can be attempted. Thickness and maximum lithic clast size measurements allow us to obtain the isopach and proximal isopleth maps as well as a median grain-size diameter ($M_d$) isoline map for the 1970 eruption. The minimum associated volume of the eruption was calculated using the isopach map defined by Baker & McReath (1975) and Smellie (2002) and the method proposed by Bonadonna & Costa (2013) using the Weibull distribution between volume and square root of isopach area. The Weibull distribution for tephra deposits is associated with less uncertainty with respect to the exponential integration, first introduced by Pyle (1989), or the power law integration (Bonadonna & Houghton 2005) when distal data are missing, as for the 1970 eruption.

Violent mafic eruptions described in the literature produce total pyroclastic deposit volumes of 0.001 to >1 km$^3$ dense rock equivalent (DRE) and dispersal areas covering 10 to >500 km$^2$ (see Wong & Larsen 2010). The 1970 eruption is characterized by an estimated total bulk volume of about 0.1 km$^3$ and a dispersal area of about 4 × 10$^3$ km$^2$ (Fig. 2), calculated using the isopachs extrapolated from the data collected at the Arturo Pratt and Bellinghausen stations and data from sea-cores in the Central Bransfield Basin reported by Fretzdorff & Smellie (2002). Those researchers found a correspondence for an ash layer to volcanic ash dating from the 1970 eruption. Deposits from the 1970 eruptions have also been reported in ice-cores from Livingston Island (Pallás et al. 2001; Geyer et al. 2006, 2008).
Following the methods proposed by Bonadonna & Costa (2013) for a median grain-size diameter (Md) isoline map from the data of Geyer et al. (2006, 2008) (Fig. 7b) and in comparison with the 1967 eruption, we can infer a column height of at least 10 km. Wind velocities for August 1970 were calculated using data from the Bellingshausen stations on King George Island (http://www.antarctica.ac.uk/met/READER/upper_air/uawind.html), with approximate values of 28–38 m s\(^{-1}\) at heights of 5500 and 9100 m, which confirm that the pyroclastic material was strongly affected by wind during deposition.

The wide dispersion of deposits could also be explained by the height of the tropopause above Antarctica, which is unusually low and in the range of 8–10 km (Weyant 1966); additionally, the stratosphere here is much less stratified than in the tropics or at mid-latitudes. This would allow for the rapid and widespread distribution of tephra in the area (Kyle et al. 1981) and would permit even small eruptions to inject material into the stratosphere. This could also explain the generally coarse grain-size character of proximal deposits, which are probably fines-depleted as shown in Figures 4 and 6c. D. PEDRAZZI

Tephrochronology studies reported by Smellie (1999) show a significant tephra record of volcanic eruptions from Deception Island in the South Shetland Archipelago (e.g. on Livingston, James Ross and Elephant Islands) showing how atmospheric wind circulation patterns at this latitude ensure a widespread distribution pattern of fall deposits.

The 1970 eruption was undoubtedly one of the most violent in the past century on Deception Island and falls into the category of violent Strombolian eruptions, similar to the eruption of Okmok volcano, Alaska (Wong & Larsen 2010) or the eruptions of Vesuvius described by Arrighi et al. (2001). It corresponds to a volcanic explosivity index (VEI) 3 of Newhall & Self (1982) and is thus of comparable magnitude to the flank eruptions of San Salvador Volcano (El Salvador) of the last 1700 years (Sofield 2004), or the eruptions of Paricutin (Mexico) between 1943 and 1952 (Luhr & Simkin 1993), or of Nabro (Eritrea) in 2011 (Smithsonian Institution 2011; Bourassa et al. 2012).

This eruption occurred in shallow seawater and onshore ice-free locations, in the same general area as the 1967 eruption but with more widely distributed vents. The 1967 episode was very similar in nature and size to the 1970 eruption and generated submarine and subaerial vents. Three overlapping pyroclastic cones (Fig. 3) with water-filled craters were created in the northwestern corner of Telefon Bay, and a further centre was located 2 km east of the new island on the shore of Port Foster between Telefon Bay and Pendulum Cove (Fig. 3). The eruption has been reported in detail by Roobol et al. (1975), who described a column height of about 2.5 km at the beginning of the eruption, rising to 6 km and then reaching a maximum height of about 10 km. The wind speed was less than 10 m s\(^{-1}\) (Valenzuela et al. 1968). As reported by Roobol et al. (1975), a thin covering of ash was noted on Livingston Island and probably on Greenwich Island as well. An ash-coated iceberg was reported close to Elephant Island, which is about 200 km to the NE of Deception Island.

On Deception Island, eruptions of high explosivity owing to magma–water interaction are common and are characterized by shallow submarine vents or are located on waterlogged shorelines or beneath the ice caps. The 1967 and 1970 eruptions demonstrate that similar explosive events are controlled differentially by the physical nature of the vent and the way in which the magma interacts with the surrounding environment, leading to several closely spaced eruptive phenomena with different characteristics and associated hazards. Although there were 42 men in the British and Chilean stations and more in the Argentine station during the 1967 eruption, no lives were lost (Roobol 1982), although some of the installations were destroyed. Despite the lack of witnesses to the 1970 eruption (Baker & McReath 1975), its characteristics, deduced from the study of its deposits, suggest a similar or even larger event compared with that of 1967, but with similar associated hazards.

The Bay of Port Foster has traditionally been used as a natural harbour by sailors and, in particular, by whalers, who installed the factory in Whaler’s Bay (Fig. 1c) that was destroyed during the 1969 eruption (Smellie 2002). Scientific bases were built by Argentina, the UK and Chile around Port Foster (Fig. 2a), the latter two being destroyed by the 1967 and 1969 eruptions (Roobol 1982). Currently, a Spanish and an Argentinean base (Fig. 2a) operate during the Antarctic summer seasons. The island has also become an important tourist attraction and was visited by almost 50000 people in 2004–2010, with annual peaks of 10000 visitors (IAATO, International Association of Antarctica Tour Operators, http://iaato.org/tourism-statistics). Thus, the volcanic risk is increasing given the high probability of an eruption in the near future.

The characteristics of the 1967 and 1970 eruptions suggest that a future eruption could have a considerable impact on the people and installations on Deception Island, particularly if it occurs during the austral summer. Therefore, a detailed reconstruction of past eruptions, as we have attempted in the present study, is a necessary first step toward a precise hazard assessment.

Conclusions

The study of the stratigraphy, lithology and depositional features of the eruption of 1970 on Deception Island allows us to reconstruct its eruptive dynamics and main physical parameters. The eruption occurred close to the area of 1967 vents, in shallow seawater and onshore ice-free locations, and led to the formation of two groups of craters, one with a cone shape and the other with maar-like craters, that reveal the differing nature of activity between two vent regions. The 1970 eruption probably had alternating magmatic and phreatomagmatic phases that deposited fallout, ballistic blocks and bombs, and caused pyroclastic surges. A total bulk volume of about 0.1 km\(^3\) has been estimated for this eruption, and its tephra deposits dispersed northeastwards as far as King George Island, >150 km away. This explosive activity is classified as violent Strombolian with VEI 3, and produced a plume that rose to at least 10 km above sea level. The 1970 eruption was similar to the 1967 one and confirms how this type of post-caldera volcanism on Deception Island is controlled by the location of the vents and magma interaction with the surrounding environment. This could give rise to closely spaced eruptions or even several phases during the same eruptive event, with different characteristics and associated hazards. A similar eruption occurring in the future on Deception Island might represent a serious hazard to the increasing number of people that visit the island during certain periods of the year and to the scientific installations that operate there.

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