# Hydrothermal Alteration on Composite Volcanoes -Mineralogy, Hyperspectral Imaging and Aeromagnetic Study of Mt Ruapehu, New Zealand

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**Abstract**

Prolonged volcanic activity can induce surface weathering and hydrothermal alteration that is a primary control on edifice instability, posing a complex hazard with its challenges to accurately forecast and mitigate. This study uses a frequently active composite volcano, Mt Ruapehu, New Zealand, to develop a conceptual model of surface weathering and hydrothermal alteration applicable to long-lived composite volcanoes. The rock samples were classified as non-altered, supergene argillic alteration, intermediate argillic alteration, and advanced argillic alteration. The first two classes have a paragenesis that is consistent with surficial infiltration and circulation of the low-temperature (40 degree C) neutral to mildly acidic fluids, inducing chemical weathering and formation of weathering rims on rock surfaces. The intermediate and advanced argillic alterations are formed from hotter (100 degree C) hydrothermal fluids with lower pH, interacting with the andesitic to dacitic host rocks. The distribution of weathering and hydrothermal alteration has been mapped with airborne hyperspectral imaging through image classification, while aeromagnetic data inversion was used to map alteration to several hundred meters depth. The joint use of hyperspectral imaging complements the geophysical methods since it can numerically identify hydrothermal alteration style. This study established a conceptual model of hydrothermal alteration history of Mt Ruapehu, exemplifying a long-lived and nested active and ancient hydrothermal system. This study highlights the need to combine mineralogical information, geophysical techniques and remote sensing to distinguish between current and ancient hydrothermal and supergene alteration systems, to indicate the most likely areas of future debris avalanche initiation.

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# Hydrothermal Alteration on Composite Volcanoes –Mineralogy, Hyperspectral Imaging and Aeromagnetic Study of Mt Ruapehu, New Zealand

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**Key Points:**

- Surface weathering and hydrothermal mineralogy was constrained using VNIR and SWIR reflectance spectroscopy and SEM-EDS analysis
- Combination of airborne hyperspectral image analysis and aeromagnetic data inversion mapped surface and buried hydrothermal alteration on Mt Ruapehu volcano
- Complex hydrothermal evolution of Mt Ruapehu is revealed using geophysical imaging techniques combined with surface alteration mineralogy

## Abstract

Prolonged volcanic activity can induce surface weathering and hydrothermal alteration that is a primary control on edifice instability, posing a complex hazard with its challenges to accurately forecast and mitigate. This study uses a frequently active composite volcano, Mt Ruapehu, New Zealand, to develop a conceptual model of surface weathering and hydrothermal alteration applicable to long-lived composite volcanoes. The rock samples were classified as non-altered, supergene argillic alteration, intermediate argillic alteration, and advanced argillic alteration. The first two classes have a paragenesis that is consistent with surficial infiltration and circulation of the low-temperature (<40 °C) neutral to mildly acidic fluids, inducing chemical weathering and formation of weathering rims on rock surfaces. The intermediate and advanced argillic alterations are formed from hotter (≥100 °C) hydrothermal fluids with lower pH, interacting with the andesitic to dacitic host rocks. The distribution of weathering and hydrothermal alteration has been mapped with airborne hyperspectral imaging through image classification, while aeromagnetic data inversion was used to map alteration to several hundred meters depth. The joint use of hyperspectral imaging complements the geophysical methods since it can numerically identify hydrothermal alteration style. This study established a conceptual model of hydrothermal alteration history of Mt Ruapehu, exemplifying a long-lived and nested active and ancient hydrothermal system. This study highlights the need to combine mineralogical information, geophysical techniques and remote sensing to distinguish between current and ancient hydrothermal and supergene alteration systems, to indicate the most likely areas of future debris avalanche initiation.



## Plain Language Summary

Groundwater heated by shallow intrusive bodies beneath an volcano can contribute heat and magmatic gasses to the groundwater. The magmatic contributions, such as sulfur dioxide, can make groundwater acidic. When the acidic groundwater rises to the surface, it chemically interacts and changes volcanic rocks, leading to hydrothermal alteration. This alteration can weaken rocks by depositing clay minerals in fractures and pore-space, and in turn inducing edifice collapse hazards. This contribution integrates mineralogy, hyperspectral remote sensing and geophysical methods to understand the hydrothermal alteration history of a seemingly unaltered volcano, Mt Ruapehu, New Zealand. Ground sampling indicates alteration mineralogy is dominated by sulfides and sulfates. Hyperspectral remote sensing measures reflected sunlight from the Earth's surface, allowing quantitative discrimination and mapping of surface minerals at high resolution using image classification. The surface alteration was jointly analyzed with the aeromagnetic inversion models to understand underground hydrothermal alteration. Aeromagnetic data is sensitive to iron-bearing, magnetic minerals that are often dissolved by hydrothermal fluids, leaving low magnetic anomalies. The combination of these methods allowed a detailed conceptualization of alteration history of Mt Ruapehu during the last 200,000 years. This conceptual model will be used to assess natural hazards associated with flank collapses using numerical models.


## 1. Introduction

Interaction of the magmatic heat and gasses ( e.g. mostly Cl, $SO_2$, $H_2O$, $CO_2$) drives infiltrating surface and circulating groundwater to ascend through the volcanic host rocks while leaching primary alkali (Na, K) and alkaline earth metals (Mg, Ca), replacing existing mineral phases and precipitating new secondary minerals at shallower depths (Ganino et al., 2019; Hynek et al., 2013; Rowe and Brantley, 1993; Rye et al., 1992). Acid-sulfate alteration forms typical intermediate and advanced argillic mineral assemblages, including phyllosilicates (Dill, 2016; Hynek et al., 2013), sulfates (Rye et al., 1992; Zimbelman et al., 2005), sulfides, and native sulfur (Inostroza et al., 2020; Piochi et al., 2015). The majority of hydrothermal alteration occurs beneath the surface in hypogene conditions; however, mineral assemblages formed under hypogene conditions are often subject to supergene weathering and erosion which can replace some of the metastable hydrothermal minerals under atmospheric pressure and temperature conditions, including dickite, anhydrite, sulfur and pyrite (Fernández-Caliani et al., 2004; John et al., 2008; Scott, 1990; Zimbelman et al., 2005). This commonly results in mineral imprinting, making interpretations of the original hydrothermal condition, fluid composition and pH challenging, particularly on complex and long-lived volcanic systems.

Hydrothermal alteration results in economic resources, such as Au, Ag, and base metal deposits; however, it also changes rock mechanical and geotechnical properties (del Potro and Hürlimann, 2009; Pola et al., 2014), promoting flank instability (Finn et al., 2001; Heap et al., 2015; John et al., 2008; López and Williams, 1993; Norini et al., 2020; Schaefer et al., 2015) and increasing the likelihood of phreatic eruptions (Mayer et al., 2017; Pardo et al., 2014). Hydrothermal alteration primarily reduces rock strength (del Potro and Hürlimann, 2009; Farquharson et al., 2019), and changes permeability within the volcanic edifice (Mordensky et al., 2019b), which may locally elevate pore-pressure promoting edifice flank instabilities (Ball et al., 2018; Collard et al., 2020; Reid, 2004). Flank instabilities, and the resulting mass flow events triggered by gravity, weather events, volcanic eruptions, magmatic intrusions, and earthquakes (Capra, 2006; Procter et al., 2014; Schaefer et al., 2018) can result in far-reaching and potentially dangerous volcanic hazards downstream from volcanoes (Finn et al., 2001).

Geological mapping and quantification of hydrothermal alteration in volcanic systems have traditionally been carried out using ground and field geological mapping, combined with Scanning Electron Microscopy (SEM), X-Ray Diffraction (XRD) and X-ray fluorescence (XRF), and isotope and fluid inclusions studies, among others, to constrain the paragenesis of the alteration mineral suites (Ball et al., 2013; Christenson and Wood, 1993; John et al., 2008; Nuñez-Hernández et al., 2020; Piochi et al., 2019; Rye, 2005; Zimbelman et al., 2005). Multispectral satellite remote sensing in the Visible and Near Infrared (VNIR – 300-1000 nm) and Shortwave Infrared (SWIR – 1000-2500 nm) is often used to upscale mapping efforts, using Landsat series (Mia and Fujimitsu, 2012; Wright et al., 2001), Advanced Spaceborne Thermal Emission and Reflection Radiometer – ASTER (Galvão et al., 2005; Rowan and Mars, 2003), and WorldView constellation (Kruse et al., 2015). Lately, hyperspectral remote sensing is becoming available for mineral alteration mapping, improving the differentiation among key indicator minerals such as alunite, jarosite, kaolinite, montmorillonite and illite, and mica, offering a cost-effective but highly sophisticated technological solution (Carrino et al., 2018; Crosta et al., 1998; Crowley et al., 2003; Hellman and Ramsey, 2004; Kereszturi et al., 2018; Swayze et al., 2014; van der Meer, 2004). However, optical remote sensing approaches are limited to quantify only the surface manifestation of hydrothermal alteration and weathering, hampering these techniques to be used for hazard assessment of geophysical mass flow.

Complementing ground and remote sensing methods, geophysical techniques, including electromagnetic resistivity, gravity and aeromagnetic surveys (Finn et al., 2018; Miller and Williams-Jones, 2016), magnetotelluric surveys (Abdallah et al., 2020; Bowles-Martinez and Schultz, 2020; Jones et al., 2008; Matsunaga et al., 2020), seismicity (Pu et al., 2020), moun imaging (Le Gonidec et al., 2019) have also been used to infer internal architecture, and locate hydrothermal fluids and zones of demagnetization of the host rock. These survey methods are often complemented with numerical methods and spring and gas chemistry data to gain insights into the thermal and chemical evolution of shallow magmatic-hydrothermal systems (Berlo et al., 2020; Collard et al., 2020; Gresse et al., 2018; Miller et al., 2020a). Therefore, geophysical inversion and numerical methods can add the depth component to quantify hydrothermal alteration processes and volumes of altered rock masses, contributing important inputs to mass flow initiation and run-out models and hazard assessment (Finn et al., 2018; Finn et al., 2001; Rosas-Carbajal et al., 2016).

Both geophysical and optical remote sensing approaches have been demonstrated to be useful for mapping hydrothermal alteration products. However, the joint use of such methods is seldomly presented on composite volcanoes to fingerprint hydrothermal alteration processes on the surface and depth. This study aims, therefore, to explore the hydrothermal alteration history of a long-lived and complex andesitic composite volcano, Mt Ruapehu, New Zealand, using a novel combination of mineralogical, hyperspectral and geophysical imaging techniques. The detailed reconstruction of the hydrothermal alteration history and conceptual model can provide inputs to predict edifice instabilities and associated geophysical mass flow hazards around composite volcanoes.

## 2. Geological Setting

The southern part of the Taupo Volcanic Zone contains Mt Tongariro and Mt Ruapehu volcanoes and several smaller inactive volcanic centers, which collectively form the Tongariro Volcanic Complex (Fig. 1A). Mt Ruapehu, 2797 m a.s.l. and 110 km$^3$, is a frequently active andesite-dacite volcano (Hackett and Houghton, 1989), formed via back-arc volcanism behind an active subduction zone at the Australian and Pacific plate boundary (e.g. Stern et al., 2006; Wallace et al., 2004).

The first deposit originated from Mt Ruapehu is dated to around 340 ky ago (Hackett and Houghton, 1989; Price et al., 2012), preserved within distal catchment areas as river aggradation terraces (e.g. Tost et al., 2015). The stratigraphic framework of Mt Ruapehu comprises four formations, on the basis of geochronology, geochemistry and stratigraphic relationships, pinpointing distinct spatial-temporal stages of volcano evolution (Fig. 1A): Te Herenga (200 to 150 ka), Wahianoa (166-80 ka), Mangawhero (50-15 ka) and Whakapapa (<15 ka) formations (Conway et al., 2016; Gamble et al., 2003; Hackett and Houghton, 1989; Price et al., 2012; Townsend et al., 2017). The only major hiatus in activity is from 80 to 50 ky, which was a period of erosion and edifice instabilities (Townsend et al., 2017).

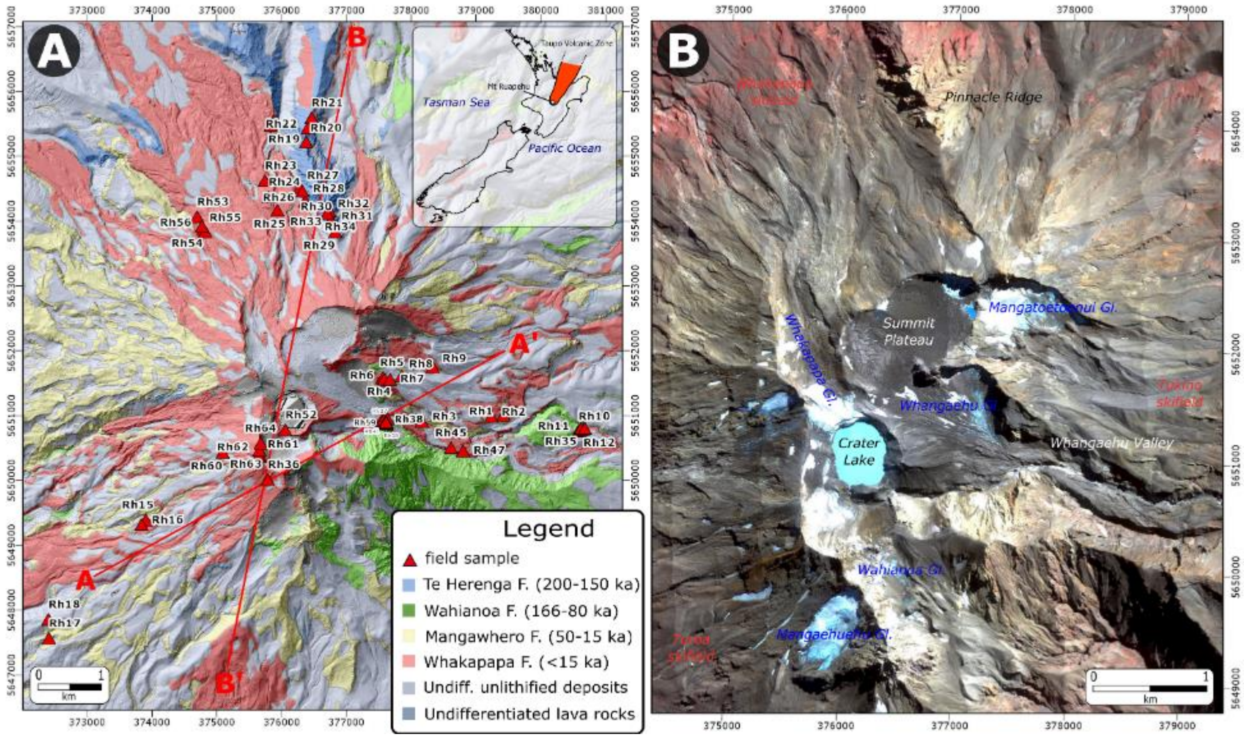



**Figure 1.** (A) A simplified geological map of Mt Ruapehu, with the major geological formations after Townsend et al. (Townsend et al., 2017), draped over a hillshade model. The red triangles show the location of the physical samples collected in this study. (B) Infrared-colored composite image of the airborne hyperspectral imagery (R = 860.1 nm; G = 650.2 nm, B = 550.7 nm) with the locations mentioned in the text.

The whole-rock chemistry of Mt Ruapehu varies between basaltic andesite and dacites (Price et al., 2012). Rock textures at Mt Ruapehu are mainly porphyritic with phenocryst abundances averaging 35-55% (Price et al., 2012); however, some lava flows exhibit aphyric texture. The main phenocrysts in decreasing abundance are plagioclase (8-39 vol%), clinopyroxene (2-14 vol%), orthopyroxene (2-10 vol%), magnetite and titanomagnetite (1-6 vol%), as well as rare occurrences of olivine and amphiboles (Conway et al., 2018; Gamble et al., 1999; Graham and Hackett, 1987; Nakagawa et al., 1999; Price et al., 2012). See detailed summary in Table S1. The composition of the plagioclase and pyroxene phenocrysts show differences within the main lithologic units, highlighting a changing magmatic source and edifice evolution over time (Conway et al., 2018; Kilgour et al., 2013).

The current volcanic edifice is made of mostly lava flows and their auto-breccias, with minor exposed volcaniclastic deposits, such as lahar, debris avalanche and tephra fall deposits (Conway et al., 2016; Townsend et al., 2017). Such lava flows seldomly reach the extensive ring plain around the volcanoes which preserved an extensive record of debris flows, debris avalanches, hyper-concentrated flows with interbedded local andesitic and distal rhyolitic tephras (Donoghue and Neall, 1996; Pardo et al., 2012). Mt Ruapehu has produced 8-10 debris avalanches in the last 340 ky. The last two debris avalanches, Murimotu (9.5 ky) and Mangaio (4.6 ky) affected the NW and E sides, and reached ca. 15 and 25 km distance, respectively (Donoghue and Neall, 2001; Palmer and Neall, 1989).

A recent period of magmatic activity occurred between September and November 1995, and June and July 1996, producing mild Strombolian to violent phreatomagmatic eruptions (Cronin et al., 1997; Johnston et al., 2000; Nakagawa et al., 1999). The 1995-96 and 2007 eruptions disrupted the top parts of the hydrothermal system, providing otherwise inaccessible samples to characterize the physico-chemical state of the host rock and hydrothermal under the currently active Crater Lake (Christenson, 2000; Christenson et al., 2010; Christenson and Wood, 1993).

A previous magnetotelluric survey data was used to model the electrical resistivity structure of the upper flanks of Mt Ruapehu and its surroundings, indicating hydrothermally altered rocks located under the Summit Plateau (Ingham et al., 2009; Jones et al., 2008). Two higher resistivity zones (20-60 Ω m) located under the Northern part of the Summit plateau at 200-500 m, and 1000-1500 m depths, are interpreted to be due to changing alteration mineralogy from smectite-illite to chlorite-rich zones (Jones et al., 2008). This region was further imaged using aeromagnetic data inversion and interpreted to be smaller demagnetized zones, corresponding to hydrothermally altered rocks (Miller et al., 2020b). Besides Crater Lake, there is no known active hydrothermal system on Mt Ruapehu; however, older parts of the Mt Ruapehu indicates extensive hydrothermal alteration history, such as the Te Herenga Formation (Mordensky et al., 2019a).

## 3. Methods and Materials

### 3.1. Field Sampling and Lab Analytics

A total of 64 samples were collected during field campaigns between 2017 and 2019 (Fig. 1A). The samples are from lava flows (interior core and brecciated exterior), and tephra deposits and range from fresh to extensively altered. Some of the samples represent subsurface lithologies (upper 200 m), brought to the surface by the 2007 eruption induced lahar and the ballistics from the 1995-96 eruption. Samples were analyzed using lab-based reflected light spectroscopy, optical microscopy, and Scanning Electron Microscope (SEM) equipped with Energy Dispersive X-ray spectroscopy (EDS).

Lab-based reflectance spectroscopy was completed using a FieldSpec 4 Hi-Res spectroradiometer, equipped with a Hi-Brightness contact probe with a sampling footprint of 10 mm in diameter. The samples were air-dried at 40 °C for 24 hours before analysis to ensure samples were dry. The spectral readings were calibrated against a white Spectralon Diffuse Reflectance Standard. Spectra measurements were completed at 2-7 “spots” on each specimen, which assured representative sampling of the variation of the alteration within each sample. Each of these “spots” included at least 100 spectral measurements. All spectral measurements were then averaged using View Spec Pro software and later exported into a spectral library. When an alteration rim/crust was present on a sample, the interior and outer rim were analyzed separately. The averaged spectral reflectance was used to recognize the typical mineral association within the samples using continuum removed spectral curves (Clark and Roush, 1984). The wavelength of the absorption features (i.e. reflectance lows) was matched manually with the USGS Spectral Library Version 7 (Kokaly et al., 2017), as well as automatically using a Spectral Feature Fitting approach implemented in ENVI (Clark et al., 1990). A list of key spectral absorption features and descriptions are summarized in Table S2.

A subset of the samples was prepared for thin section, Scanning Electron Microscope (SEM) and Energy-dispersive X-ray Spectroscopy (EDS) for petrographic and geochemical analysis. These samples were cut, mounted onto glass slides and then ground to 30 µm and polished for optical microscopy, while other samples were ground to about 100 µm, and polished for SEM-

EDS analysis. The latter batch was carbon-coated and imaged using ThermoFisher Scientific™ FEI Quanta 200 Environmental Scanning Electron Microscope operated in Back-Scattered Electron (BSE) mode under accelerating voltage of 20 kV, with a working distance of 10 mm, at the Massey University's Manawatu Imaging Centre. BSE equipped with an EDAX-EDS system was used to identify element abundance to characterize rock alteration types and secondary mineralogy.

Magnetic susceptibility measurements were made on tephra and rock outcrop and hand specimens using a Terraplus KT-10 v2 magnetic susceptibility meter. The Terraplus was held in direct contact with the surface of each sample on a flat surface if possible, measured three times, and averaged to produce one value per sample.

## 3.2. Hyperspectral Imaging Surveys

Aerial surveys were completed using a Cessna 185 survey aircraft, with two integrated imaging systems onboard, between 10:25-12:45 NZST on 31 March 2018. The survey aircraft hosted a Specim AisaFENIX hyperspectral imaging system, alongside a Nikon D810 digital single-lens reflex camera with a 35 mm lens. The aircraft was flown at 3300 feet above the ground, capturing N-S orientation survey strips with 410 m line spacing.

The GPS-tagged digital photos had a maximum ground resolution of 13.9 cm. They were captured with 60% side- and forward overlaps. This imagery was to create a structure-from-motion Digital Surface Model (James and Robson, 2012; Westoby et al., 2012). The RAW imagery was converted to JPEG to perform the initial image matching in Pix4D software package. This was followed by georectification using 3D ground control points from high-resolution aerial photographs, taken in 2016, from Land Information New Zealand, and existing 10 m terrain models captured by Horizons Regional Council. This workflow resulted in a 50 cm bare-ground Digital Terrain Model (DTM), and 50 cm an orthophoto mosaic. Both were used to process and co-register the coarser-resolution hyperspectral imagery.

The AisaFENIX hyperspectral sensor is a push-broom, full-spectrum imaging system, along with an Oxford Survey+ Global Position System and Inertial Measurement Unit (Pullanagari et al., 2016). It captures reflected light between 377 and 2500 nm with variable spectral sampling intervals of 3.3-5.7 nm for VNIR bands and of 11 nm for the SWIR bands. It has a full-width-at-half-maximum between 3.2–12.2 nm. The Field of View is 32.2°, while the Instantaneous Field of View is 0.084°. The data has 448 spectral bands with a ground resolution of 1.5 m. The raw hyperspectral sensor measurements are converted to radiance ($W \cdot sr^{-1} \cdot m^{-2}$) using sensor-specific gain and offset values, provided by Specim Ltd, Finland. The radiance imagery is corrected for atmospheric effect using ATCOR-4 algorithm, performed on the raw imaging geometry (Richter and Schläpfer, 2002). The atmospheric correction parameters were identical to Kereszturi et al. (2018). The atmospherically corrected imagery represents surface reflectance values that can be compared with spectral libraries or with other survey data. The atmospherically corrected reflectance imagery was geocoded using PARGE (Schläpfer and Richter, 2002). Since the hyperspectral imagery was captured with standard GPS positioning (±6-8 m at 2σ), there are often misalignments and shift to *X* and *Y* directions by up to 10 m. This was reduced by co-registering all image strips to the high-resolution orthophoto. The Root Mean Squared Error of the co-registration was around <2 m. The imagery was mosaiced, and then spectrally smoothed to reduce noise.

The hyperspectral imagery can be used to map the surface spatial distribution of hydrothermally altered mineral packages (Carrino et al., 2018; Murphy et al., 2015; Rogge et al., 2014; Zabcic et al., 2014), using supervised image classification. This requires training and validation data to be identified. This study uses the hydrothermal alteration mineralogy and their sample location as well as field observations to develop training and validation populations, split at 50-50% for the supervised image classification. The total area was split into 5 rock alteration classes based on the alteration mineralogy (e.g. unconsolidated-unaltered, unaltered lava rocks, supergene argillic alteration, intermediate-, and advanced argillic alteration types), and 3 general classes (e.g. water, ice/glacier and shadow). The alteration types were based on the observed mineralogy in the collected samples. This study used Random Forest classification (Breiman, 2001), which is effective at reducing overfits due to highly co-linear data (i.e. hyperspectral imagery), while also being time-efficient, and yielding accurate results (Belgiu and Drăguţ, 2016; Kereszturi et al., 2018; Pal, 2005). Random Forest algorithm constructs decision trees using a subset of the training data and variables (i.e. spectral bands). Each tree is expanded until either the maximum number of input data used or they reached the minimum impurity value (i.e. 0), based on the calculated Gini impurity measure (Belgiu and Drăguţ, 2016; Breiman, 2001). In this study, the total number of tree models was 500, and at each split of the inputs, the square-root of input total variables (i.e. 21 bands) was used. The class values were assigned using majority voting procedure based on the individual tree's prediction. The resultant classification image was assessed using independent validation population, through calculating the overall accuracy and an error matrix of user's and producer's accuracies (Liu et al., 2007).

### 3.3. Helicopter borne Aeromagnetic Survey

A helicopter-based aeromagnetic survey was carried out two weeks after the hyperspectral surveys in April 2018. We collected 800 km of magnetic data with a Geometrics G822A Ce-vapor magnetometer flown at 50 m above ground at 250 m-spaced flight lines which we reduced to 125 m over the Pinnacles and Summit Plateau. Sampling the magnetic field at 10 Hz results in a magnetic field value every 2 m. We corrected for diurnal variations using a local base magnetometer outside the survey region and subtracted the International Geomagnetic Reference Field (IGRF) at each data point location using the IGRF 2005 model. The data were levelled using widely spaced tie lies to minimize cross over difference and were gridded for visualization at 20 % of the line spacing using a minimum curvature algorithm (see Miller et al. (2020b) for full details).

To map the spatial distribution of magnetic and non-magnetic rocks within Mt Ruapehu volcano, a magnetic vector inversion algorithm was used, which is implemented in the SimPEG inversion framework package (Cockett et al., 2015; Fournier et al., 2020; Miller et al., 2020b). The inversion accounts for remanent magnetism, common in volcanic rocks, by solving for both the amplitude and direction of magnetization. The inversion returns a 3D model of apparent susceptibility that we interpret in terms of altered versus fresh rocks, where high apparent susceptibility typically reflects unaltered rocks and low apparent susceptibility increasingly altered rocks. The inversion model uses a mesh with minimum dimensions of 50×50×25 m or 25×25×10 m over the Pinnacle Ridge and Summit Plateau areas. The typical apparent susceptibility values recorded from the model range from 0 to 0.05 SI, which can be interpreted as a degree of hydrothermal alteration, from fully demagnetized/altered to magnetic/fresh rocks, respectively (Miller et al., 2020b).

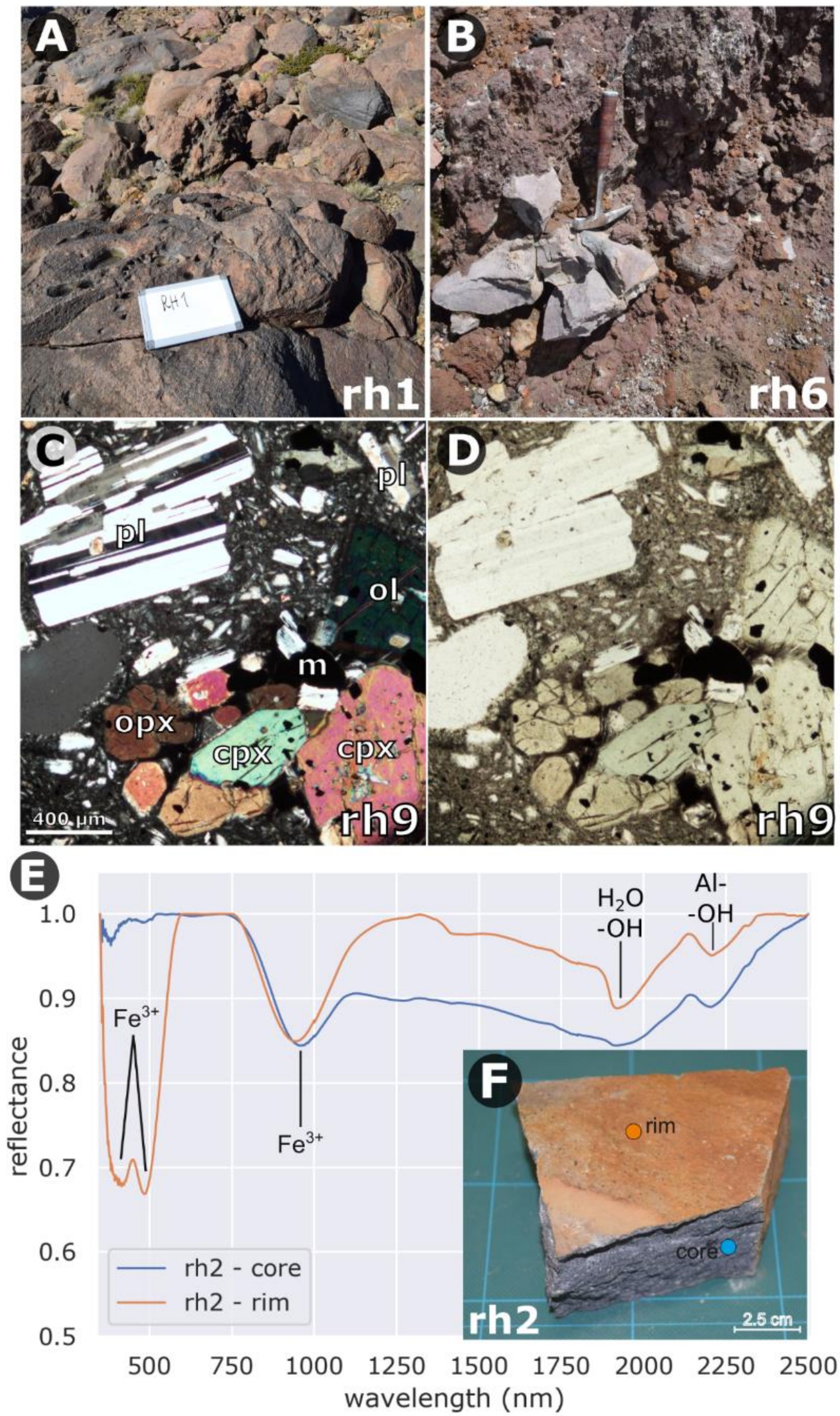
# 4. Results

## 4.1. Alteration Mineralogy

Based on the physical samples were grouped based on their alteration mineral phases mineralogy: (1) non-altered rock/deposits with minor surface weathering, (2) supergene argillic alteration with weathering rim, (3) intermediate argillic alteration and superimposed surface weathering, and (4) advanced argillic alteration. This classification scheme is based on the presence and absence of alteration minerals that represent distinct physico-chemical alteration domains (John et al., 2019; John et al., 2008; Rye et al., 1992; Zimbelman et al., 2005).

### 4.1.1. Non-altered Lithologies with minor weathering

The non-altered lithologies frequently occur on blocky lava flows with auto-brecciated horizons (Figs 2A-B) and are characterized by absorption at wavelengths of 420, 480, 500-560, 1912, 2200 nm, and occasionally ±660, ±950, ±1430 nm (e.g. rh1, rh2, rh4-9). In thin section, the studied samples show mostly fresh porphyritic, and rarely aphyric and vitrophyric textures (Fis 2C-D). The samples have about 30-50 vol% phenocrysts of plagioclase, clinopyroxene and orthopyroxene, various amounts of Ti-rich magnetite, and rarely olivine (Figs 2C-D). The phenocrysts are often euhedral to subhedral and occasionally show glomerophyric appearance. Some of the phenocrysts show chemical zonation, melt inclusions and sharp crystal boundaries, indicating their unaltered and fresh origin. The groundmass is made of tabular plagioclase microlites and volcanic glass. Some samples have a thin (≤1 mm) yellow to brown colored alteration rim, in which the groundmass is often replaced by secondary minerals, including phyllosilicates. The phenocryst phase appears to be still fresh (Figs 2C-D), but rarely Ti-magnetite crystals show trellis-type lamellae structures, especially within weathering rims. This can potentially indicate either high-temperature exsolution and oxidation during cooling (e.g. Buddington and Lindsley, 1964; Tan et al., 2016) or due to hydrothermal alteration and surface weathering (e.g. van Hinsberg et al., 2010). The measured magnetic susceptibilities of this group range from 0.002 to 0.03 SI (Miller et al., 2020b), reflecting a wide range of variability of the (Ti-) magnetite content on Mt Ruapehu (Price et al., 2012).


A
rh1
B
rh6
C
pl
pl
ol
m
opx
cpx
cpx
400 μm
rh9
D
rh9
E
reflectance
wavelength (nm)
Fe3+
Fe3+
H2O
-OH
Al-
-OH
rh2 - core
rh2 - rim
F
rim
core
rh2
2.5 cm

353

**Figure 2.** (A-B) Field photos of the fresh non-altered outcrop, and (C-D) thin section photographs of a fresh lava sample under cross-polarized light in C and plane-polarized light in D. Abbreviations: plg – plagioclase, opx – orthopyroxene, cpx – clinopyroxene, o – olivine, m – magnetite. (E) Spectral reflectance profiles of the inner core and rim of rh2 sample (F). The main elements and element-bonds are indicated the cause of the light absorption. The reflectance values have been normalized using continuum removal (Clark and Roush, 1984).

The position of the absorption in the VNIR and SWIR regions can be explained by oxidation of primary mineralogy (e.g. magnetite, pyroxene) and limited, but detectable development of secondary clay minerals (e.g. mostly in the groundmass), with and without alteration rims (e.g. Fig. 2E). Such absorption is due to the presence of hydrous phases (e.g. 1430, 1912 and 2200-2350 nm, and state transition from $Fe^{2+}$ to $Fe^{3+}$ (e.g. between 400-1000 nm) (Hunt and Ashley, 1979)). Typically, the alteration rims contain more secondary minerals, such as goethite, ±ferrihydrite, ±phyllosilicate and hematite. This observation is consistent for both tephra, breccia and lava rocks (e.g. Table S3). Moreover, most of the samples show similar spectral reflectance for both the core and rim in the SWIR region (e.g. rh19 - Tawhainui lava flow, Iwikau Member, Whakapapa formation), indicating the overall fresh, young (≤50 ky) and unaltered state of the samples.

#### 4.1.2. Supergene Argillic Alteration

These samples have additional spectral features to the unaltered lithologies, at around 650, 950, 1270, 1430, 1770, 2050, 2260, 2480 nm (e.g. rh10-11, rh21-23). These are consistent with Fe-oxides (e.g. goethite, hematite), with occasional jarosite and phyllosilicate phases, and are often limited to the alteration rim of the samples (Fig 3A). The alteration rims are often thicker than in the non-altered samples (>2 mm). The core is often comprised of fresh phenocrysts and micro-phenocryst populations characterized by sharp boundaries and sub- to euhedral crystals, lacking any pervasive alteration in the core of the samples (Figs 3C-D). This freshness of phenocrysts and micro-phenocrysts are also accompanied by magnetic susceptibilities, ranging from 0.005 to 0.02 SI (Miller et al., 2020b).

The spectroscopic data indicate the presence of both goethite (FeO(OH)) and hematite ($Fe_2O_3$) on Mt Ruapehu, with a dominance of goethite. The formation of such alteration minerals is strongly pH-dependent. Acidic (pH = 2–5) and alkaline conditions (pH = 10-14) favor the formation of goethite, while neutral pH promotes the formation of hematite (e.g. Schwertmann and Murad, 1983). However, both minerals can occur on the same hand specimen, indicating a highly heterogeneous occurrence of those minerals on a cm-scale.

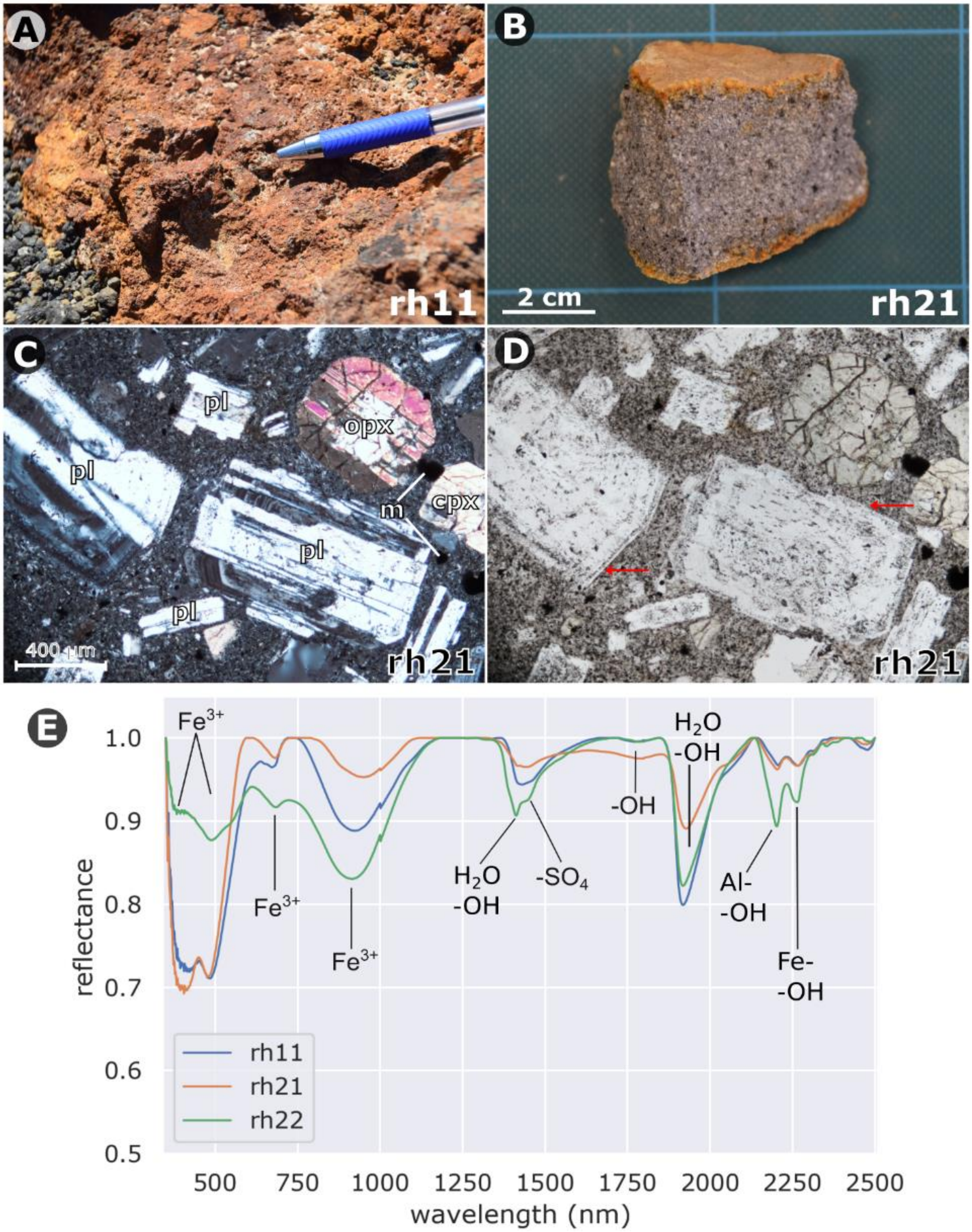



**Figure 3.** (A-B) Field and sample photos of the supergene argillic alteration showing various surface colorations and oxidations at outcrop scale. (C-D) Thin section photographs showing fresh phenocryst population with sharp contact (red arrow) viewed under cross-polarized light in

C and plane-polarized light in D. Abbreviations: plg – plagioclase, opx – orthopyroxene, cpx – clinopyroxene, m – magnetite. (E) Spectral reflectance profiles of representative samples showing supergene argillic alteration minerals.

Jarosite ($KFe_3(SO_4)_2(OH)_6$) is common in the alteration rims, suggesting a supergene origin as a weathering product of Fe-bearing rocks. It often forms from oxidation products under acidic conditions, and frequently forms in concert with ferric oxyhydroxides (Bishop and Murad, 2005). The jarosite phase (e.g. absorption feature at 2265 nm; Fig. 3E) is observed more often within the Te Herenga Formation (e.g. rh21-23) than in the Wahianoa Formations (e.g. rh10-11). This is consistent with the geochemistry of the original rocks (e.g. increased K and Fe content in the Te Herenga Formation; Table S1). The mineral assemblages are consistent with a paragenesis of a supergene argillic alteration and oxidation under atmospheric conditions, forming diverse alteration rims, depending on the primary rock geochemistry, and the exposure time to alteration processes (>50 ky). This process can be driven by metasomatism of K-bearing plagioclase phases by slightly acidic surface waters, causing surface weathering (e.g. Vasconcelos and Conroy, 2003).

#### 4.1.3. Intermediate Argillic Alteration

This group is characterized by brownish to yellowish discoloration, with a partial to full replacement of the primary rock textures (Figs 4A-C), as well as the development of pervasive alteration rims with distinct spectral reflectance (Figs 4D-E). These rocks have absorption around 380, 435, 490, 960, 1420, 1780, 1915, 2205, 2290, 2315, 2390, 2480 nm with ±650, ±2240, ±1100 nm (rh3, rh12, rh26-rh34, rh36-41). The presence of absorption features at 380, 430 and 480 nm with occasional ~650 nm and ~940 nm are due to $Fe^{3+}$ oxidation (Hunt and Ashley, 1979), indicating goethite as the main mineral phase with occasional jarosite, schwertmannite and pyrrhotite (Fig. 4E). These samples lack hematite, indicating formation under acidic conditions (Schwertmann and Murad, 1983). The SEM-EDS data further indicate that the groundmass of many alteration samples has disseminated subhedral pyrite with diameters between 5-30 µm (e.g. rh28, rh36, rh38, rh40, Figs 5A-D). Larger 300-500 µm, euhedral pyrite crystals are also present occasionally along grain boundaries, and within cavities and fractures (e.g. rh38; Fig. 5D). In addition to pyrite, Fe-rich, S-poor, occasionally Mn-rich, mineral phases have been identified, infilling cavity walls and fractures showing colloform, globular and botryoidal morphologies (Fig. 5A), is consistent with Fe-oxides (e.g. goethite), and Fe-sulfates [e.g. schwertmannite ($Fe_{16}O_{16}(SO_4)_{12\text{-}13}\cdot 10\text{-}12H_2O$)]. Jarosite occasionally appears as a pseudomorph after pyrite (e.g. rh36; Fig. 5C). The paragenesis of this phase is interpreted to be after supergene oxidation of the sulfide-rich host rock under strongly acidic conditions (e.g. Nordstrom, 1982). The SEM-EDS results show a higher volume% of pyrite in the Wahianoa than Te Herenga Formation. This can indicate time differences since those deposits are exposed to atmospheric conditions.

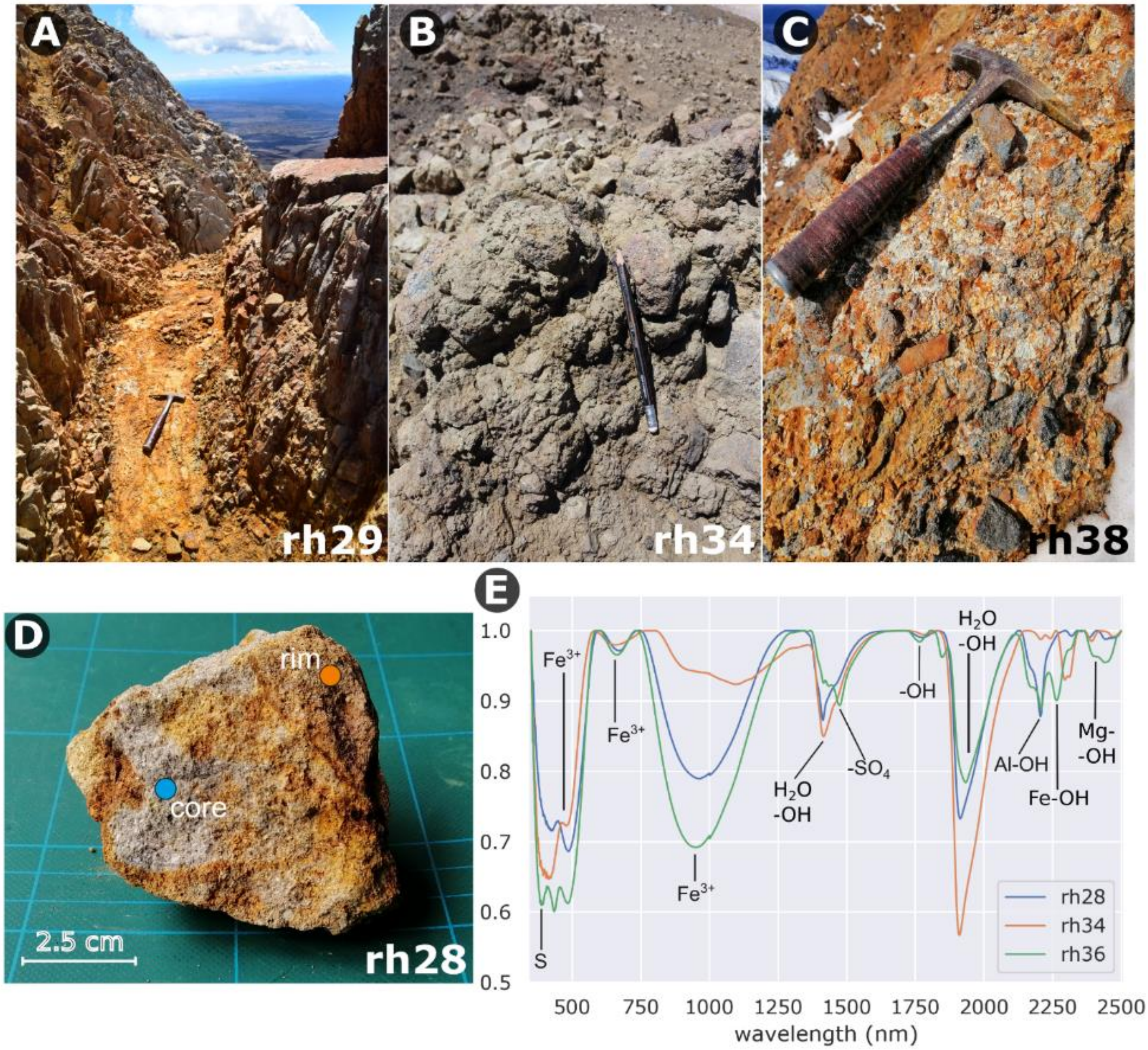



**Figure 4.** (A-C) Field photos of typical to intermediate argillic alteration on Mt Ruapehu. The samples are typically moderately to pervasively altered, showing various discoloration and clay mineral abundances. (D) Sample rh28 shows pervasive alteration preventing a clear distinction between crystal's rim and core. (E) Spectral reflectance profiles of representative samples. The spectral reflectance curves show absorption feature SWIR related to atomic vibration between Al-OH, Fe-OH and Mg-OH bonds, indicating the presence of phyllosilicates.

Based on SEM-EDS data, both primary plagioclase and pyroxene phases alter to amorphous Si-rich phases, consistent with various polymorphs of quartz. The alteration leaves abundant rim and core dissolution structures with well-developed microfractures, occasionally colloform and pseudomorph crystal habits within cavities (Fig. 5A). These are consistent with acid-induced mineral dissolution structures (e.g. Farquharson et al., 2019). Ti-magnetite phenocrysts and micro-phenocryst show commonly trellis-type lamellae textures, indicating leaching of the Fe and enrichment of Ti-oxides and silicates (e.g. rutile – $TiSO_2$, titanite – $CaTiSiO_5$), which is in association with pyrite formation. This alteration process is responsible for the decrease of magnetic susceptibility of this group, which is between 0 and 0.01 SI (Miller et al., 2020b).

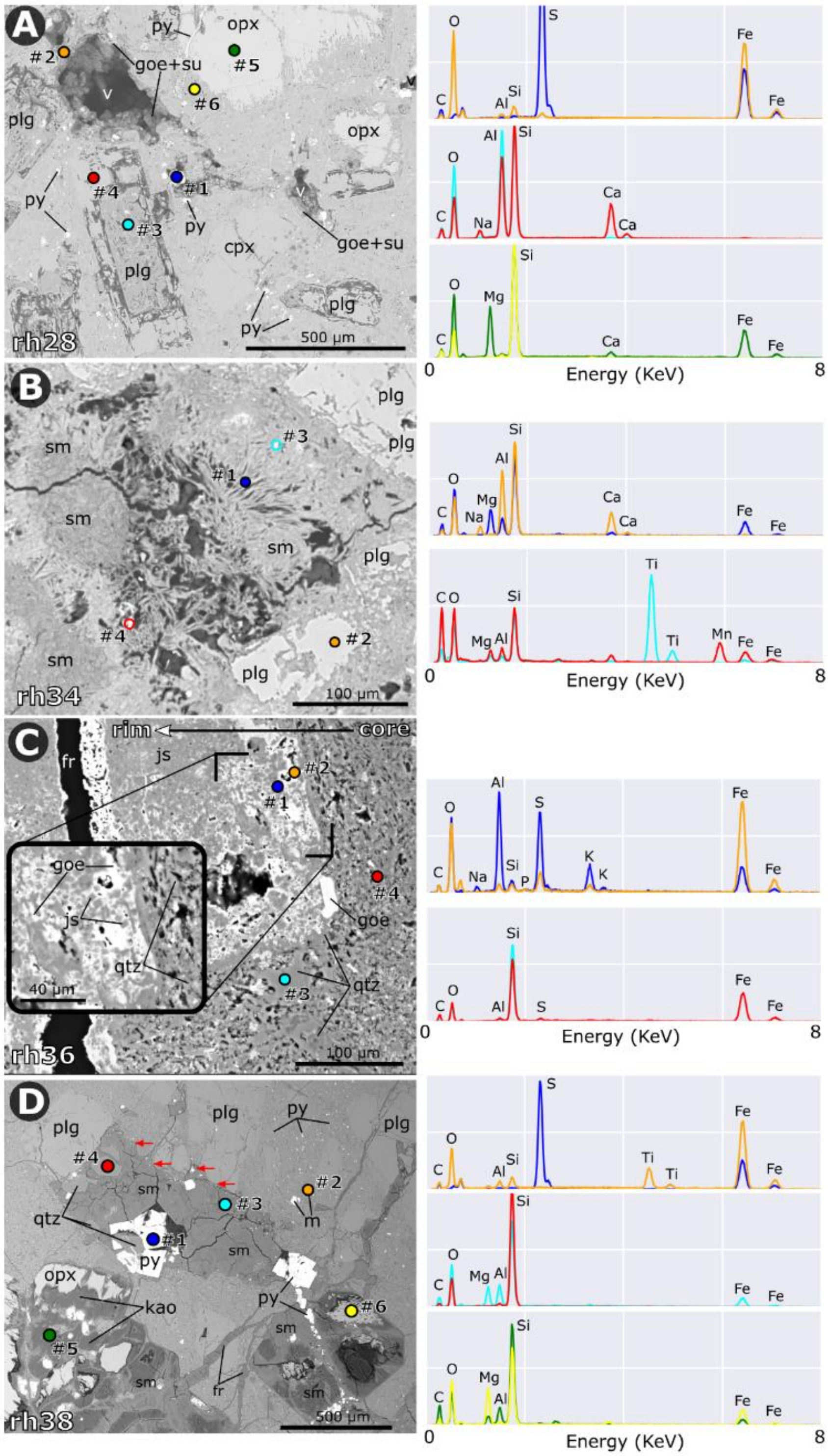


**Figure 5.** SEM (left column) and EDS (right column) results from the representative samples intermediate argillic alteration on Mt Ruapehu. The labelled EDS spots are color-coded. (A) rh28 shows typical phenocryst and micro-phenocrysts dissolution textures and extensive vug development with Fe-rich and S-poor infilling, comprising of Fe-oxides and Fe-hydroxy-sulfates. This sample is representative for the base of the Pinnacle Ridge. (B) rh34 shows vermiform, fibrous and tubular morphologies of smectite-group mineral, occupying fracture and vugs. (C) rh36 shows jarosite and goethite developed pseudomorphs after cubic pyrite. The groundmass is extensively replaced by silica (qtz). (D) rh38 shows well-developed cubic pyrite crystals co-occur with kaolin and smectite-group clay minerals (e.g. #5 and #6 on EDS) occupying vugs developed in former phenocrysts. A smaller population of pyrite occupies the interior walls of former phenocrysts, shown by red arrows. Abbreviations: v – void/vug, fr – fracture, plg – plagioclase, opx – orthopyroxene, cpx – clinopyroxene, m – titanomagnetite, sm – smectites, js – jarosite, kao – kaolinite, qtz – quartz, goe – goethite, su – Fe-sulfates, py – pyrite.

The absorption positions at 1420-1430, 1780, 2205, 2290, 2315, 2390 nm are consistent with the abundance of Na-Mg-Ca-Fe-rich phyllosilicates, including both kaolin and smectite group minerals. The distinction within phyllosilicates is challenging; however, reflectance spectroscopy in the VNIR and SWIR can detect vibrational and overtone-derived absorption features with hydrous minerals (Hunt and Ashley, 1979). This study used the shape, asymmetry and position of the absorption features to discriminate phyllosilicates and sulfates mostly in the SWIR (e.g. Table S2). Typically, samples with 1415 and 2205 nm (doublet) feature with inclination/asymmetry towards short-wavelength were interpreted as kaolin-group (e.g. kaolinite and halloysite - $Al_2(Si_2O_5)(OH)_4$). Absorption at 1415, 2205 nm and typically at 2290-2310 nm with inclination/asymmetry towards the long wavelength indicated the presence of the smectite group. The smectite group mineral were further discriminated by the position of their absorption feature at the SWIR region: montmorillonite $[(Na,Ca)_{0.33}(Al,Mg)_2(Si_4O_{10})(OH)_2\ nH_2O]$, nontronite $[(Na_{0.3}Fe_2((Si,Al)_4O_{10})(OH)_2\ nH_2O)]$, and vermiculite $[Mg_{0.7}(Mg,Fe,Al)_6(Si,Al)_8O_{20}(OH)_4\ 8H_2O]$.

The presence of phyllosilicates is often as fracture infilling minerals, showing thin crystallites with a vermiform, fibrous and tubular morphologies with hollow interiors, with Mg and Fe-enrichment (e.g. Fig. 5B). These textural and morphological features also indicate smectite group minerals (Beauchamps et al., 2019; Ece et al., 1999; Ta et al., 2017), indicating the dominance of montmorillonite and nontronite. Some samples (e.g. rh26) contains zeolite group minerals with acicular and radial crystal habits, occurring in fractures and cavity infilling.

Occasionally, euhedral to subhedral, hexagonal to prismatic, disseminated Cl-rich apatite crystals ($Ca_5(PO_4)_3Cl$) with diameters ranging from 30 to 50 μm, are embedded in phyllosilicate dominated fracture infills. Apatite is a minor, but ubiquitous mineral of plutonic and volcanic systems (Piccoli and Candela, 2002), and it is also present on Mt Ruapehu (Price et al., 2012). Constraining the paragenesis of this phase is problematic using spectroscopy, optical microscopy and SEM-EDS analysis, and it requires cathodoluminescence (Bouzari et al., 2016). The apatite found in rh34 contains a minor amount of S and Si, which can be incorporated into the apatite's crystal structure (e.g. Streck and Dilles, 1998), potentially indicating its magmatic origin.

Aluminum-Phosphate-Sulfate (APS) minerals occur in cavities as well-developed acicular crystals and are only observed in samples from the Wahianoa Formation. The SEM-EDS spectra show Ca-rich with minor peaks of Al and S, which might correspond to woodhouseite, $CaAl_3(PO_4,SO_4)(OH)_6$ (Dill, 2001; Stoffregen and Alpers, 1987). These APS minerals occasionally compliment acid-sulfate alteration in high-sulfidation epithermal systems which are rich in Ca, P, Al, dissolved from primary apatite and plagioclase phases (Imura et al., 2019; Stoffregen and Alpers, 1987).

The mineral associations, their spectral, chemical and textural characteristics are all consistent with intermediate argillic alteration, formed in a highly acidic environment by hydrothermal fluids with temperatures between 150-250 °C (John et al., 2019; Simmons et al., 2005). This resulted in abundant pyrite as an oxidation product of the ascending $H_2S$-rich fluids below and around the paleo water table within the proto-Ruapehu edifice, within the Te Herenga and Wahianoa Formations. The pyrite-bearing rocks were then subject to supergene alteration after exposure to the atmospheric condition following flank collapses (Palmer and Neall, 1989; Tost et al., 2015) and erosion (e.g. glaciation and fluvial activity). This led to the formation of smectite-group minerals, such as montmorillonite and nontronite, and Fe-sulfates, potentially schwertmannite, both filling in fractures and vugs. Smectite group minerals often form after the initial oxidation of pyrite, producing acid water, followed by hydrolysis of feldspar and acid water buffering by the host rock, and colloidal deposition in open cavities and fractures at low-temperature (≤40 °C) (Fernández-Caliani et al., 2004). This paragenesis often produces minor barite ($BaSO_4$) phase (Fernández-Caliani et al., 2004), which has been identified as a minor phase in XRD (Mordensky et al., 2019a).

#### 4.1.4. Advanced Argillic Alteration with minor silicification

Advanced argillic alteration occurs in vertical cliffs, and as reworked volcaniclastics deposits along the upper Whangaehu Valley, around the present-day Crater Lake area, and sampled as ballistic blocks from past eruptions (Figs 6A-C). These samples are often characterized by white, grey to pale yellow colors with abundance native sulfur, gypsum, amorphous silica precipitation and pyrite crystals with diameters up to 2 mm (Figs 6A-D). This group shows typical VNIR-SWIR absorption features around 380, 430, 480, 940, 1412, 1920, 2170, 2205, 2450 nm with ±390, ±1430, ±1475, ±1490, ±1783, ±2240, ±2260, ±2320, ±2400 nm (Fig. 6E). The spectral absorption position is consistent with goethite, phyllosilicates with the dominance of kaolin- group minerals over the smectite group, native sulfur, sulfates and sulfides. The SEM imagery shows completely dissolved phenocrysts and micro-phenocrysts of all primary minerals (e.g. Fig. 7A) The groundmass is often completely replaced by amorphous quartz (e.g. with localized silicification; Figs 6A and 7). Occasionally, samples show intense local silicification with complete to partial replacement of groundmass and phenocrysts, accompanied by minor sulfur and barite precipitation (e.g. blocks from the shore of the Crater Lake – rh52). The Ti-magnetite phases of the primary volcanic rocks have been completed altered, leaving Ti-rich residue and abundant pyrite crystals forming euhedral to subhedral on the outside of former magnetite crystals, and disseminated as a groundmass (Figs 7A and C). The destruction of magnetite produces very low magnetic susceptibility (e.g. 0 to 0.001 SI).

The groundmass occasionally has stockwork textures and colloform banding with vein and veinlets filled by native sulfur (e.g. rh44; Fig. 7B) and anhydrite (e.g. rh45; Fig. 7C). Anhydrite ($CaSO_4$) can be formed by progressive removal of acid anion species, such as $SO_4$, from the hydrothermal fluids, or by the exchange of aqueous H+ with cations in the host rock (Smith et al., 2017; Zimbelman et al., 2005). This paragenesis can also result in precipitation of kaolinite (Hynek et al., 2013). Samples exposed on the surface for an extended period were subject to depletion of their sulfur content (e.g. rh12) and hydration processes, forming gypsum, $CaSO_4 \cdot 2H_2O$ (e.g. rh48).

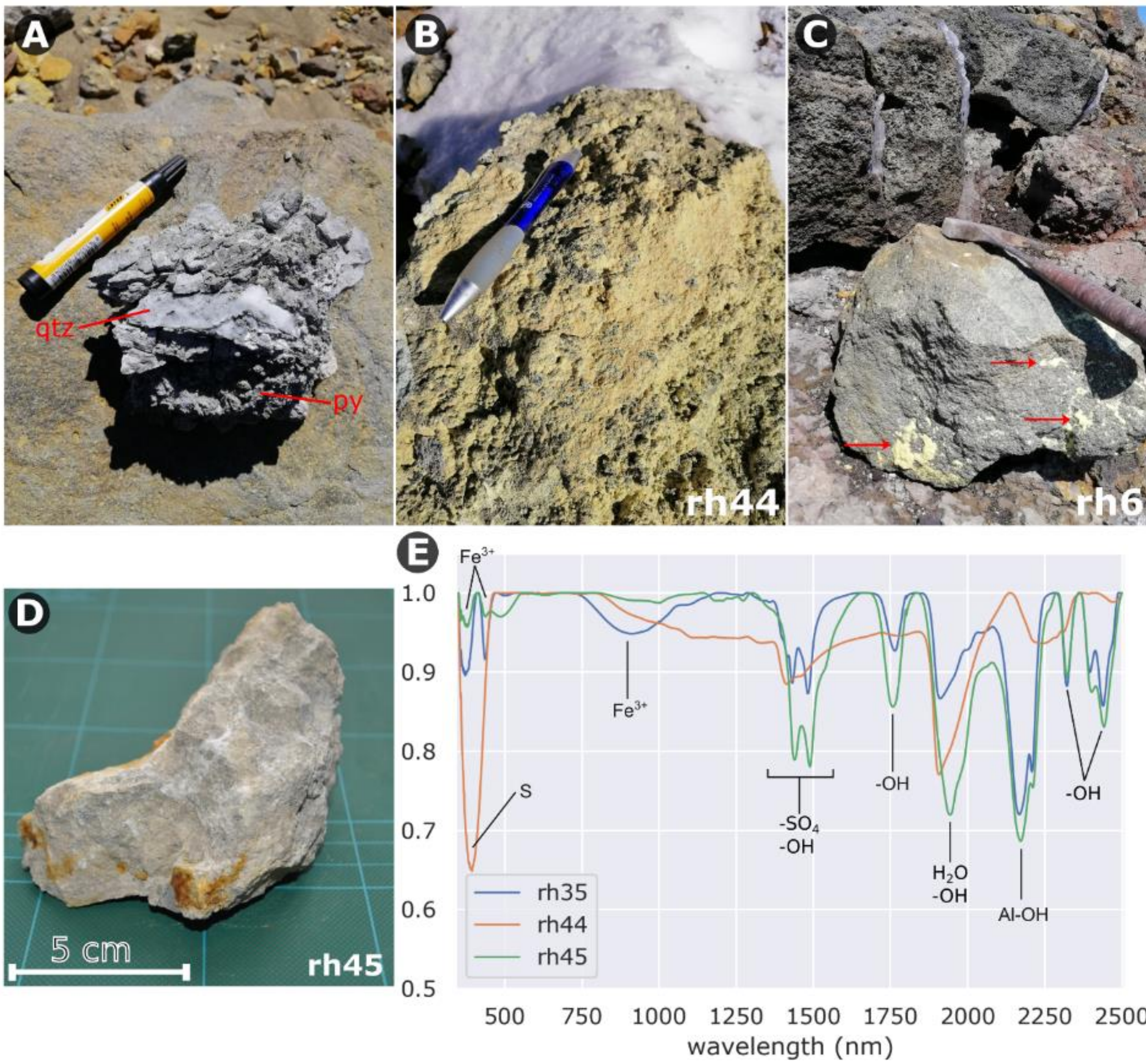



**Figure 6.** (A-D) Hand specimen samples of the advanced argillic alteration on Mt Ruapehu. The samples are often disintegrated, white to yellow in color, and show visible precipitation of pyrite/marcasite and native sulfur. (E) Spectral reflectance profiles of representative samples. The spectral reflectance curves show typical absorption features at around 1400-1500 nm which is associated with sulfates ($-SO_4$), while strong absorption due to Al-OH bonds and presence of crystalline water, indicating alunite and phyllosilicates (e.g. kaolin group minerals)

Besides anhydrite and gypsum, this group has alunite ($KAl_3(SO_4)_2(OH)_6$). Alunite exhibits spectral absorption features (e.g. Fig. 6E) due to the vibrations of hydroxyl (-OH) and metal-oxygen bonds (Al-OH) and lattice vibrations (Bishop and Murad, 2005). The alunite occurs on Mt Ruapehu as tabular crystals and as a constituent of the groundmass (Figs 7A and C). Both reflectance and SEM-EDS data indicate that alunite on Mt Ruapehu is not a pure endmember, but they show both K and Na enrichments. The origin of alunite is due to the oxidation of acidic fluids (e.g. $H_2S$) between the groundwater table and the surface (Zimbelman et al., 2005), or as supergene alteration of sulfides (Bladh, 1982). Currently, there is no active hydrothermal manifestation (e.g.

fumaroles, hot springs) besides the vent-hosted hydrothermal system beneath Crater Lake. The current Crater Lake and its hydrothermal system precipitate Na-alunite as oxidation of ascending $H_2S$ hydrothermal fluids (Christenson and Wood, 1993). The formation of alunite within the Wahianoa Formation is interpreted to be formed under hypogene conditions due to its current stratigraphic position (e.g. exposed only at the lower parts of the Whangaehu valley). However, further isotope and radiometric dating are needed to confirm its relationship with the currently active hydrothermal system under the Crater Lake area.

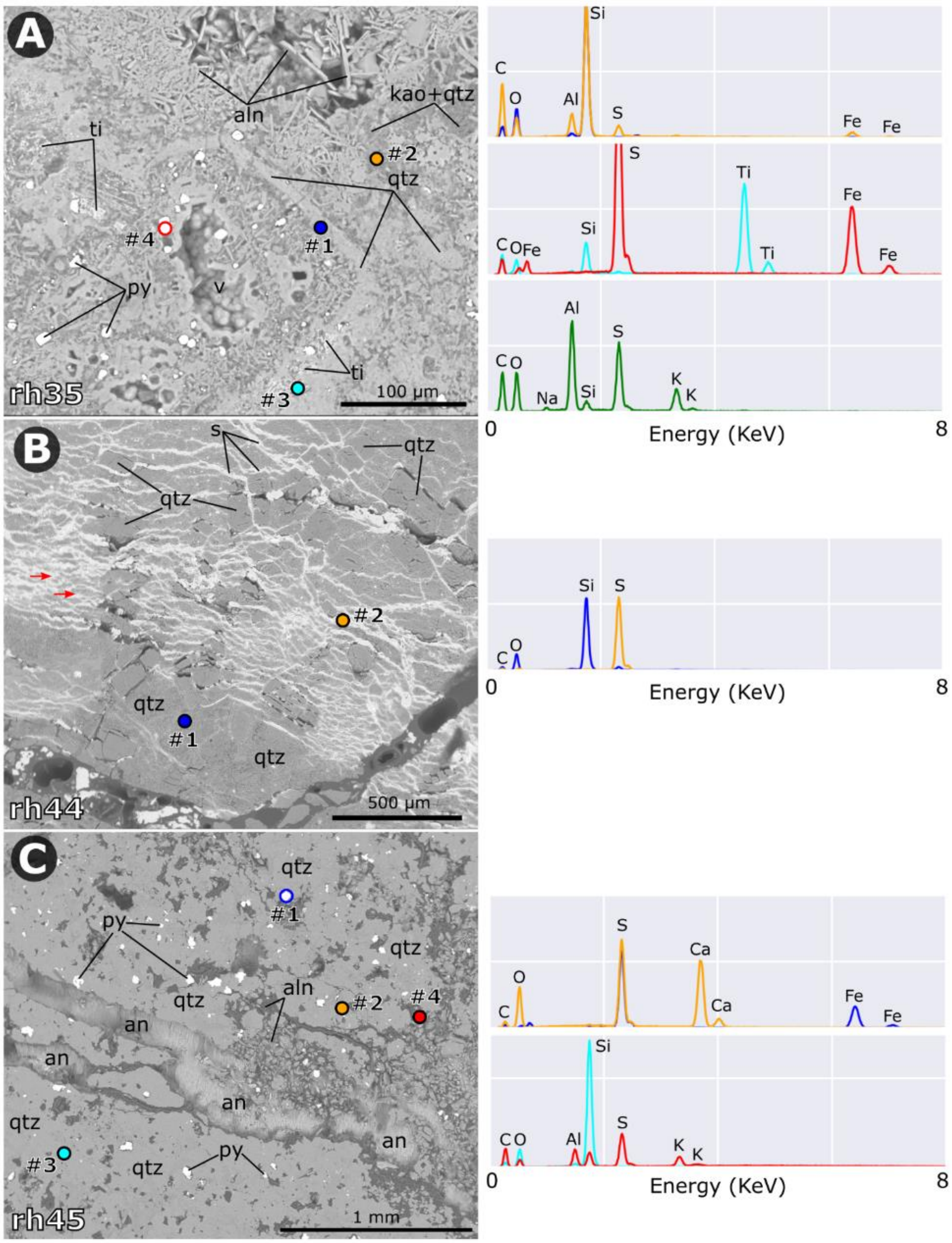



**Figure 7.** SEM (left column) and EDS (right column) results from the representative samples advanced argillic alteration on Mt Ruapehu. The labelled EDS spots are color coded. (A) rh35

shows tabular crystals of alunite, occurring together with kaolinite, pyrite and quartz. The groundmass and phenocryst are completely replaced by the alteration mineralogy; however, scattered Ti-rich phases are residues after the primary Ti-rich magnetite population (ti). (B) rh44 show extensive veinlets of precipitated native sulfur. (C) rh45 shows anhydrite filled fracture surrounded disseminated pyrite, alunite and quartz. Abbreviations: v – void/vug, kao – kaolinite, qtz – quartz, py – pyrite, al – alunite, s – sulfur, ti – Ti-rich phase.

The mineral associations are consistent with advanced argillic alteration (e.g. Sillitoe and Hedenquist, 2003) formed from a low pH of 1-4, hot hydrothermal fluids (120-300 °C), circulating within a magmatic-hydrothermal system (Boyce et al., 2007; John et al., 2019; John et al., 2008; Simmons et al., 2005; Swayze et al., 2014). However, the hydrothermal alteration can change on a small-scale between advanced argillic and intermediate argillic alteration styles.

### 4.2. Surface Mapping using Hyperspectral Imaging

The airborne hyperspectral image was used to create a hydrothermal alteration map of Mt Ruapehu through supervised image classification using a Random Forest algorithm (Fig. 8). The training process for the image classification was guided by the hydrothermal alteration mineralogy from SEM-EDS and spectroscopy analysis. The image classification accuracy using an independent validation population is 92.6%. Full retrieval of alteration classification on the surface is affected by the heavy cover of tephra, snow and ice on the surface of Mt Ruapehu (35.7% of the total area; Table 1) and the presence of shadows (Fig. 8A). The second largest unit mapped is the unaltered lava rocks (22.8 $km^2$). Supergene, intermediate and advanced argillic alteration have much smaller spatial extent on the surface, composing 4.8 $km^2$, 2.4 $km^2$ and 0.2 $km^2$ of the total area, respectively (Table 2). Accounting for the extensive surface cover, these hydrothermal alteration zones are expected to be minimum figures. The spectral average of the input training data shows distinct differences, including spectral features at 405, 493, 670 and 995 nm (goethite, hematite), 1160 nm (smectites), 1430-1495 nm (kaolinite, alunite, jarosite), 1763 nm (alunite), 2174 nm (alunite), 2200-2210 (Al-rich phyllosilicates), 2265 nm (jarosite), 2300-2390 nm (Fe- and Mg-rich phyllosilicates) (Fig. 9). These indicate that those minerals are critical to spectrally separate the mapped hydrothermal alteration types using airborne hyperspectral imagery.

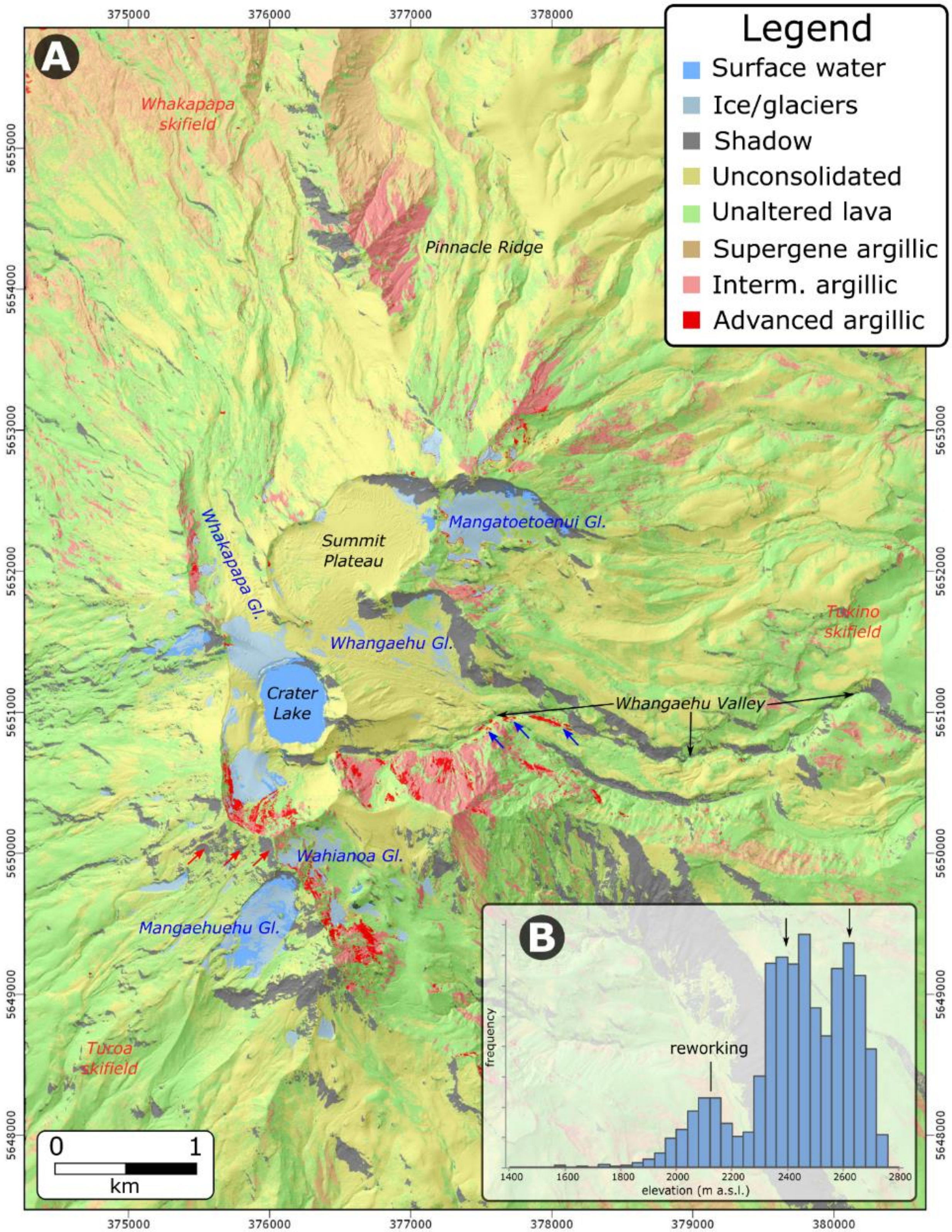



**Figure 8.** Hydrothermal alteration map of Mt Ruapehu based on the airborne hyperspectral imagery (A). The blue arrows show the location of the reworked deposits of the advanced argillic

alteration (e.g. samples rh44, rh45) along the Upper Whangaehu valley. Red arrows indicate areas with outcropping intermediate alteration (e.g. rh36), misclassified as shadow. The inset (B) shows the elevation histogram of the advanced argillic alteration rocks mapped by hyperspectral imaging, showing they occur predominantly at two distinct elevations (black arrows).

The error matrix shows numerous misclassifications between unconsolidated tephra and unaltered lava rocks (Table 1). This can be due to the similar chemical composition, resulting in similar reflectance profiles. Therefore, the spectral discrimination is most likely due to overall intensity of the reflected light, which, in turn, is a function of the grain size and illumination geometry (e.g. Clark and Roush, 1984). Another misclassification occurs between the supergene argillic and the intermediate argillic alteration types (Table 1). However, this misclassification occurred less frequently due to the presence of an absorption feature at 671 nm (e.g. goethite, ferrihydrite and hematite).

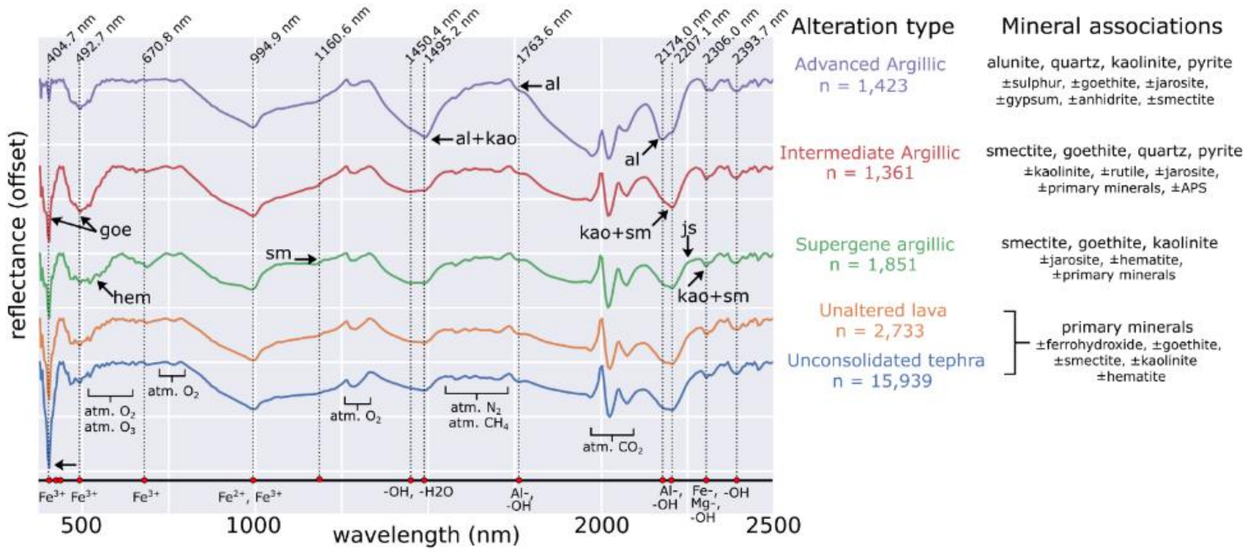



**Figure 9:** Continuum-removed spectral reflectance curves for the alteration classes on Mt Ruapehu, showing spectrally distinct absorption features of each class and the inferred mineralogy (black arrows) due to their unique mineralogy (right column). The graph also indicates the location of the atmospheric gasses, manifesting as noise in the hyperspectral data. The spectral curves are offset for clarity. Abbreviations: kao – kaolinite, al – alunite, sm – smectites, js – jarosite, goe – goethite, hem – hematite.

The most spectrally distinct class is the advanced argillic, due to its unique and pronounced absorption located at 2174 nm (alunite) and 2205 nm (kaolinite) (Fig. 9). The mapped occurrences correspond to three distinct elevation regions (Fig. 8B): (1) The lower occurrences correspond to the reworked deposits, along with the Whangaehu valley (blue arrows in Fig. 8A). The other two populations can indicate a spatial and temporal difference between the formation of advanced argillic suites (e.g. Wahianoa Formation versus the current Crater Lake hydrothermal system).

**Table 1:** Error matrix of the supervised image classification using the independent validation data (vertical columns) against the image classification results (horizontal rows). The values are pixel numbers. The highlighted values highlight the misclassifications.

| | | Image class | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | classes | water | ice | shade | tephra | unaltered lava | supergene argillic | intermediate argillic | advanced argillic | Total |
| Validation class | water | **3,381** | 132 | 44 | 0 | 0 | 0 | 0 | 0 | 3,557 |
| | ice | 0 | **1,952** | 0 | 0 | 0 | 0 | 0 | 0 | 1,952 |
| | shade | 0 | 0 | **5,820** | 4 | 0 | 0 | 0 | 0 | 5,824 |
| | tephra | 2 | 0 | 0 | **15,599** | 699 | 9 | 3 | 18 | 16,330 |
| | unaltered lava | 5 | 0 | 0 | 1,094 | **1,819** | 77 | 69 | 0 | 3,064 |
| | supergene argillic | 0 | 0 | 0 | 47 | 26 | **1,203** | 42 | 0 | 1,318 |
| | intermediate argillic | 0 | 0 | 0 | 0 | 68 | 148 | **727** | 2 | 945 |
| | advanced argillic | 0 | 0 | 0 | 0 | 0 | 0 | 3 | **595** | 598 |
| | Total | 3,388 | 2,084 | 5,864 | 16,744 | 2,612 | 1,437 | 844 | 615 | 33,588 |

**Table 2:** Area statistics and image classification accuracy by image classes and hydrothermal alteration types on Mt Ruapehu.

| **Classes** | **Alteration type/surface feature** | **User's Accuracy [%]** | **Producer's Accuracy [%]** | **Total area [pixel]** | **Total area [km$^2$]** |
|---|---|---|---|---|---|
| **class 1** | surface water | 95.05 | 99.79 | 202,329 | 0.46 |
| **class 2** | glaciers/snow | 100 | 93.67 | 402,495 | 0.91 |
| **class 3** | shadow | 99.93 | 99.25 | 1,420,459 | 3.20 |
| **class 4** | unconsolidated tephra | 95.52 | 93.16 | 8,601,991 | 19.35 |
| **class 5** | non-altered lava | 59.37 | 69.64 | 10,158,097 | 22.86 |
| **class 6** | supergene argillic alteration | 91.27 | 83.72 | 2,125,567 | 4.78 |
| **class 7** | intermediate argillic alteration | 76.93 | 86.14 | 1,075,363 | 2.42 |
| **class 8** | advanced argillic alteration | 99.5 | 96.75 | 93,699 | 0.21 |

### 4.3. Subsurface Mapping using Aeromagnetic Data

The 3D depth of hydrothermal alteration can be mapped using aeromagnetic data and subsequent inversion models (Finn et al., 2018; Finn et al., 2001; Miller et al., 2020b), if the lack of the magnetic susceptibility is due to hydrothermal alteration which dissolves (Ti-) magnetite crystals. Based on the SEM imaging, there is a systematic dissolution of (Ti-) magnetite with increasingly pervasive hydrothermal alteration (Fig 10). Therefore the magnetic susceptibility measurements and the inversion model of Mt Ruapehu (Miller et al., 2020b) indicates the hydrothermal alteration, which can be used to link surface alteration patterns from the hyperspectral image classification (Figs 10D and 11). However, identifying low magnetic susceptibility caused by hydrothermal alteration is complicated by the variable amounts of magnetite content and its Ti-impurity in the host rock (Fig. 10D), meaning that some fresh, low magnetite content andesites can have similar magnetic susceptibility to supergene alteration. Intermediate and advanced argillic alteration deposits tend to have, however, generally lower magnetic susceptibility (e.g. ≤0.01 SI; Fig 10D).

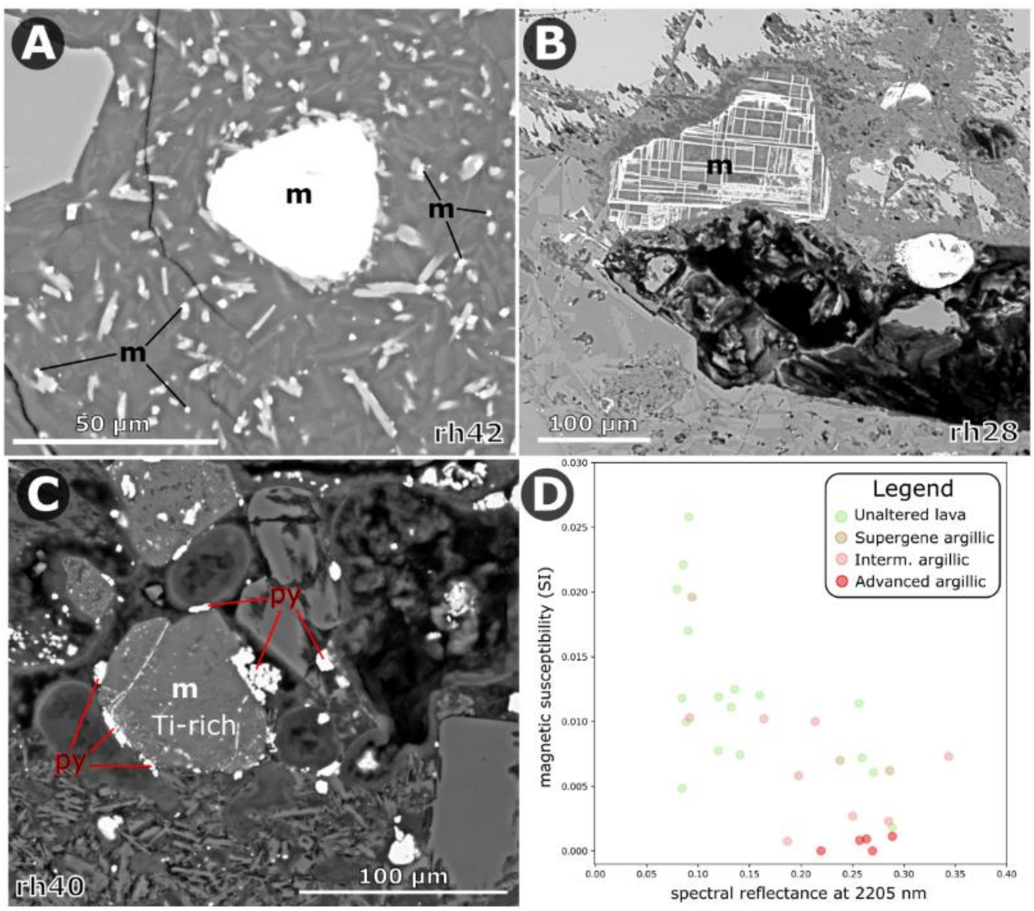



**Figure 10.** Titanomagnetite dissolution textures as a function of hydrothermal alteration types: (A) non-altered, (B) intermediate with well-developed trellis structures, and (C) advanced argillic alterations. (D) The graph shows the relationship between measured magnetic susceptibility values (SI) and spectral reflectance at 2205 nm.

The largest region of hydrothermally altered rocks is located within the Pinnacle Ridge on the southeastern flanks and to a moderate extent beneath the entire Summit Plateau. The cross-cutting dykes exposed at the Te Herenga Formation show higher magnetic susceptibility compared to the lavas of the Wahianoa Formation, possibly due to the larger grain size of the intruded rocks. The low to moderate levels of magnetic susceptibility (e.g. Turoa side; Fig. 11) are due to volcaniclastic deposits, mixed with glacial sediments and thin intercalated lava flows (Conway et al., 2016; Townsend et al., 2017), which some of might also be hydrothermally altered (Miller et al., 2020b). The demagnetized zones beneath Summit Plateau are patchy (Fig. 11B), in agreement with electrical resistivity highs observed in the magnetotelluric model, which suggest the dominance of chlorite (Jones et al., 2008). The geophysical model linked with alteration mineralogy and surface alteration distribution from the hyperspectral imaging indicates that the hydrothermal alteration interpreted in the aeromagnetic model mostly corresponds to intermediate argillic alteration. Intermediate argillic alteration is rich in smectite-group minerals, which can explain the observed the high electrical resistivity zone beneath the northern part of the Summit Plateau [e.g. “R1” anomaly in Jones et al. (2008)]. Moreover, this alteration type has a wide range of magnetic susceptibilities (0.001 to 0.01 SI) (e.g. Fig. 10D), contributing to the patch-work patterns in the aeromagnetic data (Miller et al., 2020b).

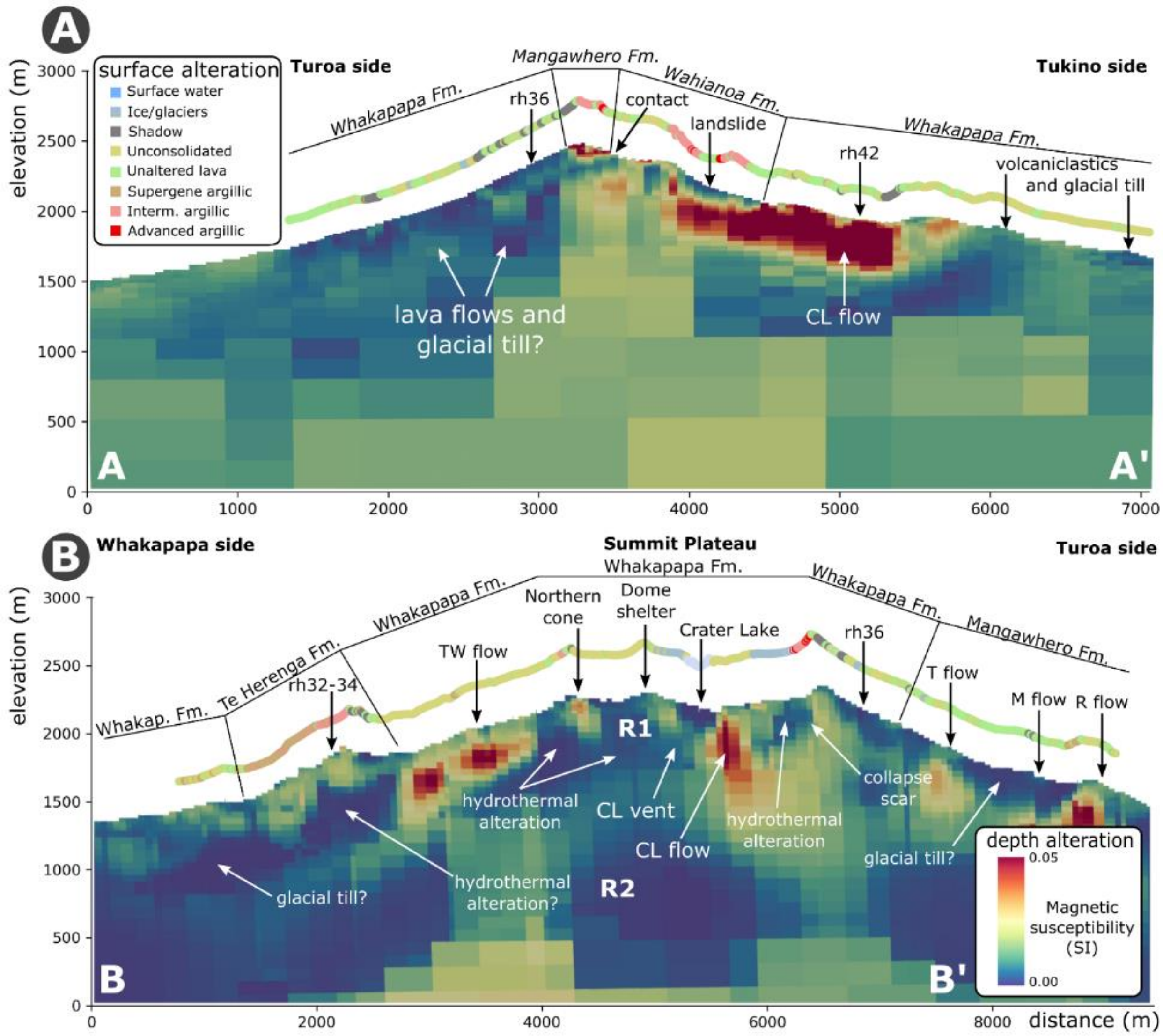



**Figure 11.** Cross-section through the magnetic susceptibility model of Mt Ruapehu along the A-A' and B-B' profiles from Fig. 1. On (B), R1 and R2 correspond to the electric resistivity highs from Jones et al. (2008). The dots above the topography show the hydrothermal alteration types mapped on the surface using hyperspectral imaging. The colors correspond to Fig. 11, and the profiles are offset by 200 m for clarity. Abbreviations: CL vent – Crater Lake vent; CL flow - Crater Lake Member lava of Whakapapa Formation (5-y); TW flow - Iwikau Member of Whakapapa Formation - Tawhainui flows (6 ky); T flow - Turoa Member of Whakapapa Formation (15-12 ky); M Flow - Mangaehuehu Member of Mangawhero Formation (45-42 ky); R flow - Rangataua Member lava of Whakapapa Formation (10-15 ky).

## 5. Discussion

### 5.1. Integration of Hyperspectral Imaging with Airborne Geophysics

Airborne hyperspectral remote sensing can identify hydrothermal alteration minerals, and through image classification, a hydrothermal alteration map can be created. The hydrothermal alteration map only represents surface alteration, which can be heavily hampered by the surface cover (e.g. tephra, glacial till, alluvium and colluvium accumulation). However, the distribution of the hydrothermal alteration can still be reconstructed with the aid of ancillary field data through microscopic petrographical and mineralogical information. Hyperspectral imaging provides a versatile framework for hydrothermal alteration mapping, which correlates well with the magnetic susceptibility model of Mt Ruapehu (e.g. Fig. 11). Conversely, the inversion of aeromagnetic data can indicate spatial distribution of subsurface alteration at depth, but it provides no information on the style of hydrothermal alteration, and interpretation of moderate alteration can be misled by unaltered rocks with similar magnetic susceptibility. Furthermore, aeromagnetic data is also insensitive to rock water saturation that is a key determinant in slope stability (Finn et al., 2018; Miller et al., 2020a). Thus, limitations in both hyperspectral remote sensing and geophysical data can be overcome by integrating both methods, allowing a comprehensive assessment of both surface and subsurface alteration patterns.

The paragenesis of hydrothermal alteration types and volumes of the altered rock masses provide a great conceptual model for volcanic hydrothermal systems, vastly improving our understanding of volcanic evolution and associated natural hazards. Specifically, at Mt Ruapehu, since the hydrothermal alteration is only exposed along ridge tops and on steep slopes off of these ridges because of tephra cover, the extent of hydrothermal alteration is seemingly minor on the surface. However, geophysical data (e.g. aeromagnetic) suggests a much larger extent is altered. This is likely the case at similar long-lived volcanoes worldwide. This calls for the need to integrate remote sensing and geophysical datasets for developing new volcano assessment tools to monitor and map shallow hydrothermal alteration within composite volcanoes.

Further direction to utilize hyperspectral remote sensing for hydrothermal alteration mapping can include mapping of individual mineral species through spectral feature matching (Clark et al., 2003), or wavelength mapping (van der Meer et al., 2018). Moreover, alteration minerals often occur as an intimate granular mixture, requiring spectral unmixing algorithms (Roberts et al., 1998), or using regression approaches with synthetically mixed training data (Okujeni et al., 2013). These research directions will need dedicated studies in the future.

## 5.2. The Role of Mineral Imprinting

The weathering and hydrothermal alteration on Mt Ruapehu has produced diverse mineral suites, some of which can be formed through multiple paragenesis including supergene modification through oxidation of sulfides, such as pyrite exposed to groundwater or surface water (Fig. 12A). These processes cause changes to meta-stable hydrothermal alteration minerals, leading to "imprinting", complicating the reconstruction of hydrothermal alteration history on long-lived composite volcanoes. For example, supergene oxidation can form jarosite (in Fe-rich low pH conditions), alunite (in moderate pH and Al-rich conditions), as well as hydroxy-sulfate and oxyhydroxide minerals (Bishop and Murad, 2005; Schwertmann and Murad, 1983; Zolotov and Shock, 2005), which cannot be discriminated by spectroscopic techniques and hyperspectral remote sensing. Mt Ruapehu has abundant Fe-oxides, mostly goethite, which can be formed from both oxidation processes of primary mineral phases rich in Fe (e.g. titano-magnetite, clino- and orthopyroxenes) and oxidation and breakdown of hydrothermal pyrite (Brady et al., 1986; Noack et al., 1993). The abundance of goethite, and its confinement to higher flanks, can also be attributed to the extensive glacial history of Mt Ruapehu. The chemical weathering occurring beneath glaciers is primarily driven by microorganic activity, thriving on meltwater solution rich in oxidized pyrite, silica and anions, such as $SO_4^{2-}$ (Mitchell et al., 2013; Rutledge et al., 2018).

The integration of field sampling, hyperspectral and aeromagnetic data have, however, allowed us to constrain the spatial distribution of inactive and currently active hydrothermal systems, allowing discrimination of mineral imprinting processes on Mt Ruapehu (Fig. 12B). These correspond to the oldest Te Herenga Formation, outcropping at the Pinnacle Ridge (Fig. 1, 8), in which the intrusion-related hydrothermal system lead to acid sulfate alteration, causing primary mineralogy to be altered to pyrite, phyllosilicates, quartz-dominated alteration minerals, superimposed with extensive smectite formation due to prolong surface oxidation of pyrite (Fig. 12B). This leads to distinct vertical and stratigraphic changes in a sulfide-to-clay ratio within the altered rocks. Areas <2000 m a.s.l. show less extensive oxidation of pyrite (e.g. rh28) than the elevated, >2100 m a.s.l parts of the Pinnacle Ridge (rh32-34). This can be explained by the delay in surface erosion and thus the exposure of the sulfides to atmospheric $O_2$, leading to enrichment of smectites.

The second oldest part of Mt Ruapehu is the Wahianoa Formation that hosts extensive areas of intermediate argillic alteration (Fig. 12B). The alteration mineralogy and textures (e.g. sharp boundaries of the pyrite crystals) are better preserved than within the Te Herenga Formation, indicating much shorter exposure to atmospheric conditions. This observation is in line with the geological history of the upper Whangaehu valley (Fig. 1 and 8), which developed after the Mangaio flank collapse 4.6 ky ago (Donoghue and Neall, 2001). Furthermore, the Wahianoa Formation also hosts advanced argillic alteration (e.g. alunite, pyrite, quartz and phyllosilicates). This can originate from (1) surface oxidation of pyrite, (2) within the magmatic-hydrothermal system that existed during the Wahianoa Formation time, or (3) a steam-heated overprinting alteration due to the proximity of the currently active Crater Lake hydrothermal systems by condensation of magmatic vapor into the later outflow of groundwater (Fig. 12B). In (2) and (3), the alunite formed directly due to the ascent-driven oxidation of $H_2S$-rich fluids and wall-rock interactions. Alunite, however, occurs with anhydrite on Mt Ruapehu, which is more consistent with the paragenesis of (2). The distribution of alunite bearing rocks, therefore, can indicate the position of the paleo water table within the Wahianoa eruptive center between 80-50 ky ago. The lowest levels of in-situ alunite occurrence on Mt Ruapehu is at 2250 m a.s.l. This elevation is just slightly higher than the maximum elevation of the Pinnacle Ridge (2237 m a.s.l.), potentially indicating the role of erosion on the exposure of alunite-bearing rocks.

### 5.3. A Model for Hydrothermal Alteration

Aeromagnetic data observes that Mt Ruapehu currently has an active but spatially confined vent-hosted hydrothermal system (Christenson and Wood, 1993). This is in sharp contrast with the neighboring Tongariro Volcanic Complex, in which has abundant hydrothermal surface manifestation, including fumaroles and hot pools (Miller et al., 2018; Moore and Brock, 1981). This observation is in line with the hyperspectral image-derived hydrothermal alteration map, which does not indicate pervasive hydrothermal alteration across the summit area. While there is limited surface hydrothermal alteration on Mt Ruapehu (i.e. intermediate and advanced argillic styles), covering 2.6 $km^2$ or only 5% of the total area (Fig. 8), the location of this alteration is scattered throughout the mapped areas. This indicates a complex spatial-temporal evolution of hydrothermal activity at Mt Ruapehu that reflects discrete development stages over the last 200 ky (Fig. 12).

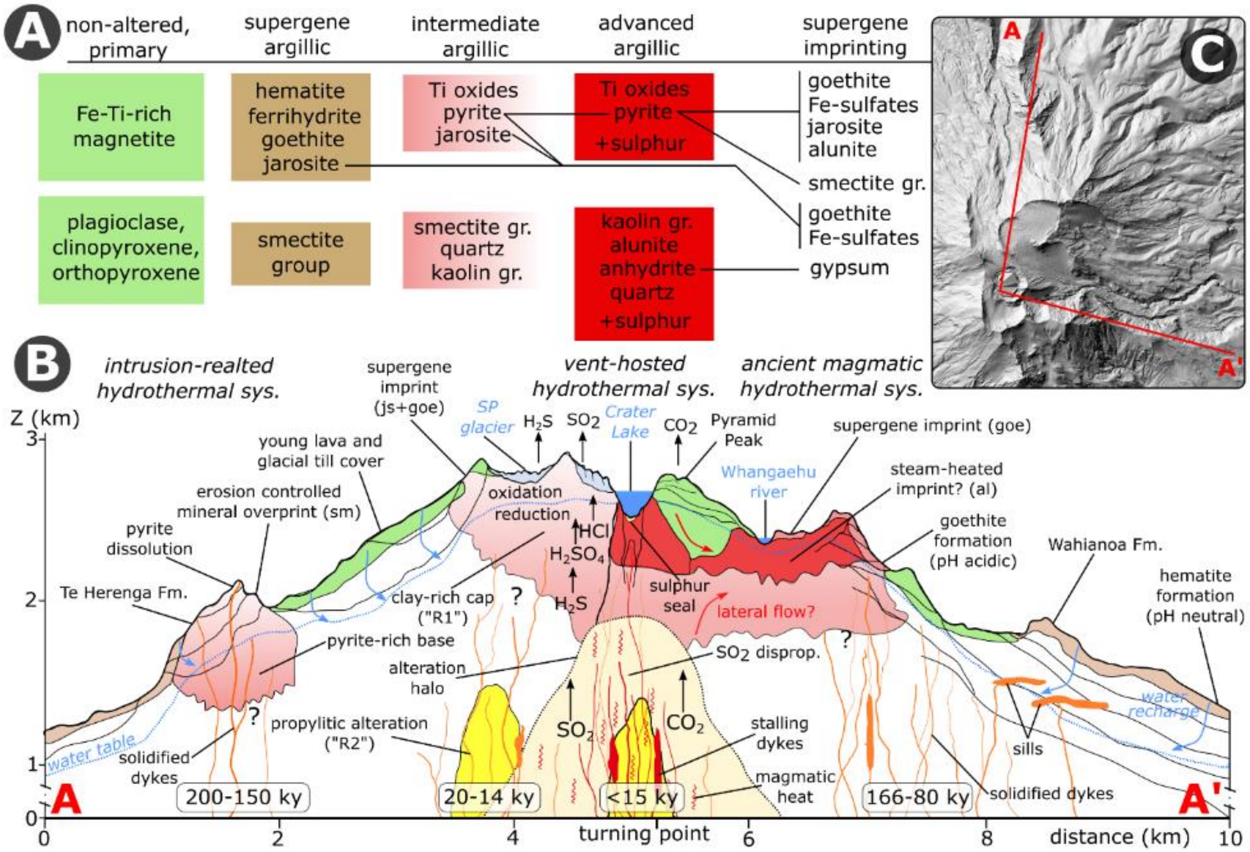



**Figure 12.** A model for surface weathering and hydrothermal alteration on Mt Ruapehu volcano. (A) Hydrothermal alteration sequence of typical primary minerals (green), and their supergene argillic (light brown), intermediate (light pink) and advanced (red), their supergene imprinting mineral associations. (B) The distribution of hydrothermal alteration types and the main hydrothermal features on Mt Ruapehu along the A-A' profile line in (C). Abbreviations: SP glacier – Summit Plateau glacier.

At Mt Ruapehu, the source of the magmatic heat and gasses are located at depths between 2 and 9 km as a network of stalling and crystallizing dykes, forming a melt-rich crystal mush inferred from volatile content and major element chemistry of groundmass glass and phenocryst-hosted melt inclusions (e.g. Kilgour et al., 2013). This zone has also been imaged as a resistivity anomaly in magnetotelluric (Ingham et al., 2009) and seismic tomography surveys (Rowlands et al., 2005). This network of magma releases magmatic gasses (e.g. $H_2O$, $CO_2$ and $SO_2$; Fig. 12B), which after disproportionation can form acidic hydrothermal fluids dominated by $H_2S$, and precipitate native sulfur (Christenson et al., 2010; Mavrogenes and Blundy, 2017). This type of hydrothermal fluid promotes acid sulfate wall-rock alteration within the volcanic edifice above the magmatic heat source (Rye et al., 1992; Zimbelman et al., 2005). The low pH hydrothermal fluids then induce wall-rock alteration at depth underneath the currently active Crater Lake. A similar ancient hydrothermal system has been imaged by magnetotelluric surveys beneath the northern Summit Plateau as a vertically elongated high resistivity zone, which can correspond to a chlorite-rich altered zone (Jones et al., 2008). Chlorite is a typical indicator mineral of propylitic alteration (Neal et al., 2018). A broader region around this zone also appears to be demagnetized (e.g. Fig. 12B; Miller et al., 2020b), which can correspond to the development of an enlarged alteration halo (Fig. 12B).

Propylitic alteration often grades vertically into intermediate argillic alteration (e.g. John et al., 2019), which outcrops only sporadically around the Summit Plateau, as indicated by the hyperspectral remote sensing data (Fig 8). This type of alteration occurs as the ascending $H_2S$-rich hydrothermal fluids react with the andesitic host rock's ferrous minerals (e.g. titano-magnetite) to form sulfides, such as pyrite, under reducing conditions (e.g. rh38, rh40, rh28, rh33; Fig. 12A), as observed similarly at Mt Rainer, USA (John et al., 2008). This leads to the formation of disseminated pyrite crystals. On Mt Ruapehu, well-developed, cubic pyrite crystals are typical, indicating hydrothermal fluid supersaturation conditions and higher formation temperatures of ~250 °C (e.g. Murowchick and Barnes, 1987). The dissolution of Ti-magnetite population is also responsible for demagnetization of the host rock, making it possible to detect hydrothermal alteration using aeromagnetic techniques (e.g. Fig. 11B; Miller et al., 2020b). This zone has also been mapped as a higher-resistivity zone on magnetotelluric data (e.g. R1 in Fig. 13B; Jones et al., 2008), indicating the dominance of intermediate argillic alteration beneath the entire Summit Plateau (Fig. 12B). This area corresponds to the relict hydrothermal system of the Paretetaitonga and Tureiti cones, active between 20 and 12 ky ago (Townsend et al., 2017). Besides the precipitation of pyrite, and enrichment of Ti, the wall-rock alteration on Mt Ruapehu also leads to K and partial Na and Ca depletions, and Si, Al, Fe, Mg and O enrichments, along with the formation of kaolin and smectite group minerals (Fig. 12A).

At locations of intense hydrothermal fluid circulation, the intermediate argillic alteration transitions into advanced argillic alteration, characterized by the abundance of sulfate minerals (e.g. alunite), and localized zones of silicification (e.g. vuggy texture). Sulfates (e.g. alunite, anhydrite and barite) often form around or above the water table within a hydrothermal system due to oxidation of the ascending $H_2S$-rich fluids (Fig. 12B) (Rye, 2005; Zimbelman et al., 2005). These are limited to areas directly beneath the Crater Lake and within the Wahianoa Formation (Fig.12B). The origin of sulfate precipitation within the Wahianoa Formation can be either (1) associated with the magmatic-hydrothermal system developed during the Wahianoa Formation, or (2) formed as a steam-heated alteration on the margin of the currently active Crater Lake hydrothermal systems due to lateral flow of groundwater (Fig. 12B). The origin can be further investigated using stable isotope geochemistry and K-Ar radiometric dating on the K-phase within the alunite.

### 5.4. Volcanic Hazard Implications

The influence of alteration on physical and mechanical rock properties depends on rock type, initial microstructural and physical properties of the rock, nature of the hydrothermal fluid, and duration of rock-fluid interaction, resulting in varying processes of mineral dissolution, replacement, and/or precipitation (Heap et al., 2020 and references therein). This study described the production of kaolin and smectite phyllosilicate clay minerals as a result of acid-sulfate alteration. The main consequences of the formation of these clay minerals are the replacement of strong rock by a weaker, clay-rich rock, precipitation of clay into pores and microfractures, and the deposition of clay along bedding structures or at the interface between volcanic deposits (Heap et al., 2019; Watters et al., 1995). The degree/intensity of clay alteration combined with the degree of saturation has been shown to alter overall strength, density, and elasticity of volcanic rock (Mordensky et al., 2019a; Siratovich et al., 2016; Watters et al., 1995; Wyering et al., 2014). Experimental observations have also shown that an increase in alteration increases the propensity for pore collapse or ductile failure behavior (Mordensky et al., 2019a; Siratovich et al., 2016), which decreases porosity and permeability as the material is deformed and pores are compacted (e.g. Farquharson et al., 2017). This, in turn, has implications for pore fluid pressure, magma degassing, and eruption characteristics (Cassidy et al., 2018; Heap et al., 2019; Okumura and Sasaki, 2014). For example, a reduction in permeability from clay precipitation into pores and fractures can prevent the migration of hydrothermal fluids or magma degassing, leading to increased pore pressures that can decrease material strength (leading to collapse events) and increase the likelihood of phreatic eruptions (e.g. Day, 1996). The presence of clays can also reduce the effective pressure required for ductile behavior, resulting in an anomalously shallow ductile zone that may prevent brittle fracturing and thus volcano-tectonic seismicity prior to some volcanic eruptions (Mordensky et al., 2019a).

A thorough investigation of the type and distribution of alteration is required to highlight potential source areas for collapse and mass flows events (e.g. debris avalanches and landslides). The existence of a single vent-hosted hydrothermal system at Mt. Ruapehu suggests that current (ongoing) hydrothermal alteration may be limited to the active summit crater and conduit (Fig. 11B). However, spectroscopy and aeromagnetic data suggest an abundance of alteration exists in the upper parts of the south and southeast flanks, either due to lateral fluid migration and/or older/inactive hydrothermal systems. The mineralogical, hyperspectral and geophysical data all indicate widespread hydrothermal altered rocks underneath the Summit Plateau, which is likely to be due to intermediate argillic alteration (Fig. 14B). This alteration is abundant in smectite group minerals that can precipitate into pores and fractures, thus decreasing rock permeability and preventing the migration of hydrothermal fluids beneath the Summit Plateau. The reconstructed mineral suites and hydrothermal alteration map can be combined with ongoing rock mechanical studies at Mt Ruapehu to provide an improved understanding of the geomechanical properties and rock mass behavior. This, combined with volume estimates of altered bodies from aeromagnetic data (Miller et al., 2020b), is vital information for both defining the probability of failures initiating from these areas and implementing new numerical simulations of mass flows from composite volcanoes.

## 6. Summary and Conclusions

The present study integrates alteration mineralogy with hyperspectral imaging and aeromagnetic data inversion to provide a conceptual model for hydrothermal alteration. These geophysical and geochemical techniques can also essential to map the 3D distribution of altered rock masses on composite volcanoes. However, the methods used in this study separately would not be able to fully describe the hydrothermal alteration. A combined approach is therefore promoted here to better understand such complex geosystems. This is particularly important for composite volcanoes with long-lived and nested eruptive centres, which can leave behind relict hydrothermal systems, potentially invisible on the surface.

Hyperspectral imaging is extremely effective at delineating surface hydrothermal alteration through image classification techniques, allowing quantitative recognition of hydrothermal alteration minerals. Furthermore, airborne imaging system allows very high-resolution imagery to be acquired, benefitting a detailed spatial mapping of hydrothermal alteration styles due to their unique combination of mineralogy. However, this technology currently is not capable of assessing deep-seated hydrothermal alteration and was limited by unconsolidated tephra, snow and ice cover, and shadow.

Aeromagnetic inversion models can complement airborne hyperspectral imaging to quantify subsurface hydrothermal alteration, by mapping the volume of demagnetized rocks. Hence, the combination of the airborne remote sensing and geophysical approaches allows the creation of a detailed, three-dimensional conceptual model for supergene weathering and hydrothermal alteration processes over the last 200 ky at the Mt Ruapehu volcano.

Water saturation can be an important driven for pore pressure fluctuation in volcanic aquifers. It cannot be detected by aeromagnetic surveys, nor with hyperspectral remote sensing. A useful add-on can be for future studies to combined those survey techniques above with electrical and electromagnetic geophysical methods, such as Direct-Current resistivity, induced polarization and airborne electromagnetics for including groundwater levels water saturation. These methods need to be included in current monitoring methods to improve our understanding of the trigger mechanism of slope failures.

The detailed surface and subsurface imaging data provide here versatile and high-resolution baseline information for future studies, which is critical to assess future volcanic activity and spatial/temporal changes on frequently active volcanoes. The methods used herein can be extended to other composite volcanoes worldwide. Results from these studies, in combination with mechanical information on rock-strength, can be applied to identify the volume and configuration of structurally weak material. Better identification and delineation of flank instability hazards, obtained through the application of the alteration models, can be used to improve hazard assessment and mitigation efforts around active volcanoes.

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The airborne hyperspectral image is publicly available at: 10.5281/zenodo.3905352

The aeromagnetic data is publicly available at https://figshare.com/articles/Data_and_python_script/11289149

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