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Geological mapping strategy using visible near-infrared–shortwave infrared hyperspectral remote sensing: Application to the Oman ophiolite (Sumail Massif)

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[1] An airborne hyperspectral survey of the Oman ophiolite (Sumail Massif) has been conducted using the HyMap airborne imaging spectrometer with associated field measurements (GER 3700). An ASD FieldSpec3 spectrometer was also used in order to constrain the spectral signatures of the principal lithologies cropping out in the surveyed area. Our objective was to identify and map the various igneous lithologies by a direct comparison at high spectral resolution between field and airborne spectra despite strong variations in outcropping conditions such as (1) lighting, (2) surface roughness geometry, (3) blocks coated with red/brown patina and exfoliation products, or (4) deep hydrothermal weathering. On the basis of spectral signatures, we are able to distinguish three end-members of olivine-orthopyroxene bearing assemblages in the mantle sequence: (1) harzburgites, (2) dunites, and (3) a harzburgite with interstitial carbonate. Because plagioclase is spectrally featureless in the wavelength range studied it cannot be detected. In the crustal sequence, we therefore identified four end-members with variable abundance of clinopyroxene: (1) massive gabbros, (2) amphibolized (upper) gabbros associated with intrusive dykes, (3) wehrlite with high serpentine content, and (4) gabbro-norite (a lithology not previously recognized in the studied area). With the exception of wehrlite, spectra of olivine-rich end-members display characteristic Mg-OH narrow absorption features caused by their high serpentine content. We take advantage of this observation to split the data into two subsets, corresponding to the mantle and crustal sequences, respectively. Pixels of an image often correspond to heterogeneous areas in the field and a direct comparison between airborne and in situ spectra is not straightforward. However, comparing spectra of pixels associated with the most homogeneous areas in the field with the spectra acquired in situ at the same location, we found a systematic change both in mean intensity and overall spectral shape. Dividing each spectrum by its low-pass trend removes the effects caused by surface light scattering associated with each scale of analysis and results in an exceptional match between field and airborne spectra. However, the

albedo information is lost and as a consequence, rock types only characterized by albedo change cannot be discriminated. A spectrum of a mixture of powdered minerals is usually seen as a linear combination of mineral spectra proportional to their abundance. However, this is no longer the case when minerals occur in complex arrangements in rock types. We thus develop a synthetic spectral library of all possible combinations of rock types covering the surface area of a pixel and use a simple distance calculation to identify the best match between each pixel and modeled spectra. This procedure allows the determination of the fractional cover of each rock type in a given pixel and to establish maps for each spectral end-member. The final product is a geological map, derived from the combination of end-member fractional cover maps, and is broadly consistent with the existing geological maps. Beyond this general agreement which demonstrates the potential of this new approach for geological mapping, imaging spectrometry allows (1) to map in detail the outline of the Moho north of Maqsad and (2) to identify a new crustal sequence enriched in silica south of Muqzah, revealing the presence of orthopyroxene, the nature and distribution of which are of relevance to the petrological and tectonic understanding of the Oman ophiolite evolution.

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1. Introduction

[2] The Oman ophiolite (Figure 1a) is one of the largest (about 30,000 km²) and best-preserved fragments of fossil oceanic lithosphere presently exposed at the Earth's surface. It is composed of a wide diversity of ultramafic to mafic rocks. The precise chemical composition, distribution, and contact relationships of these rocks, together with the nature of their alteration products reflect physicochemical (magmatic, hydrothermal, and weathering) processes that prevailed during their genesis at an oceanic spreading center, their thrusting, and their final exposure as an ophiolite on the Arabian margin.

[3] An accurate geological map of the Oman ophiolite is needed for advanced petrologic and tectonic studies. Lithological mapping of the Oman ophiolite was accomplished at the scale of the entire Oman Mountains (e.g., at 1:500,000 by *Glennie et al.* [1974]), at a regional scale (e.g., at 1:100,000 by both the Open University and Bureau de Recherches Géologiques et Minières (French Geological Survey) (BRGM) teams during the eighties) or more locally at a massif scale (e.g., at

1:40,000 by *Amri et al.* [1996]). The task was not easy for two main reasons:

[4] 1. Although some rock types are different in morphology and color, other rock types may look quite similar in outcrop (and thus have to be grouped in the same unit on the geological map).

[5] 2. Whatever the quality and resolution of aerial photographs, all “classical” geological maps imply a significant part of extrapolation between a restricted number of ground control points.

[6] Consequently, errors, omissions, and simplifications often lead to misinterpretations or at least simplified geological interpretations. Remote sensing can then be an effective technique to map extended geological features, enhanced by excellent outcropping conditions due to the arid climate of Oman [e.g., see *Pontual*, 1990].

[7] It has been shown that using spectrometry, subtle lithological variations can be detected based on both electronic and vibrational processes related to primary and secondary minerals [e.g., *Hunt*, 1977]. Since fresh rocks are rarely exposed at the Earth's surface, geologists pay particular attention to the 2.10–2.50 μm wavelength range where

Figure 1. (a) Location of the main Oman ophiolite Massif and of the studied area covered in Figure 1b. (b) Outline of the HyMap survey over the generalized geologic map covering part of the Sumail Massif. This synthesis was developed by compiling and standardizing the 1:100,000 scale maps of Nakhl [Rabu *et al.*, 1986], Fanjah [Villey *et al.*, 1986], Birkat Al Mawz [Hutin *et al.*, 1986], and Samad [de Gramont *et al.*, 1986]. The projection system employed is the UTM 40N, Datum WGS 84. Dashed frames indicate both restricted sectors studied by Combe *et al.* [2006].

absorption features of weathering minerals are well expressed [e.g., Pinet *et al.*, 2006a]. Applications using multispectral satellite imagery (less than 10 channels) were largely exploited in Oman to map the ophiolite. Using Landsat Thematic Mapper (TM), which has only one channel between 2.10 and 2.50 μm , Rothery [1987] and Abrams *et al.* [1988] could distinguish dyke-rich parts of the crustal sequence and map harzburgites and dunites present in the mantle sequence. However, various types of mafic-ultramafic rocks characteristic of the crustal section could not be well identified. Later, Denniss *et al.* [1994] discriminated mantle rocks

and separated gabbros from wehrlites using the JERS-1 OPS instrument (three channels in the 2.10–2.50 μm spectral range). Both approaches identified large-scale structures and lithological boundaries using simple composite images. However, improvements of the geological maps were limited by the small number of channels of these broad band sensors.

[8] Progress in remote sensing instruments such as imaging spectrometers [Goetz *et al.*, 1985], Airborne Visible InfraRed Imaging Spectrometer (AVIRIS) [Vane *et al.*, 1984], or HyMap: Hyper-

Figure 2. Simplified 3-D diagram summarizing the main units forming both the mantle and the crustal units of the Sumail Ophiolite sequence [after *Juteau et al.*, 1988]. Main lithologies: (1) Harzburgite; (2) cumulate and residual dunites; (3) layered gabbro; (4) planar-laminated gabbro; (5) recrystallized isotropic gabbro; (6) high-level isotropic gabbro; (7) sheeted dyke complex; (8) lower extrusive lavas; (9) undifferentiated wehrlitic intrusive bodies; (10) differentiate intrusive bodies; (11) upper extrusive lavas.

spectral Mapper (HYMAP) [Cocks *et al.*, 1998] for example, allows one to record a continuous spectrum with hundreds of narrow channels for each pixel. Target identification using this technique is generally based on the analysis of the shape of the diagnostic absorption features specific to surface composition (e.g., band wavelength position, depth, width, and shoulders). This technique has already been applied to other peridotite complexes [e.g., *Chabrilat et al.*, 2000; *Launeau et al.*, 2002, 2004]. In Oman, based on a recent HyMap survey, earlier investigations have been undertaken by *Pinet et al.* [2005, 2006b] and *Combe et al.* [2006] to identify and map mineralogical units. Focusing on two specific areas of the Sumail Massif (less than 70 km², Figure 1b), *Combe et al.* [2006] compared airborne spectra to spectra associated with pixels carefully chosen, considering them as lithological end-members. However, regarding a spectrum as being representative of a distinct lithology at the pixel scale of the airborne image is a major assumption which must be cautiously assessed since the 6 × 6 m surface cell associated with the instrument resolution is generally composed of mixed lithologies and various surface states.

[9] The purpose of this study is to map the main ophiolite lithologies covering a larger area of the Sumail Massif (i.e., ~1380 km², solid black line in Figure 1b), relying on a straightforward comparison between field rock-type spectra and image pixel spectra, as close as possible to the approach followed for standard geological mapping.

[10] We first describe the spectral signature of each lithology encountered in the field. Then we detail the methodology developed for both comparing spectra acquired at different scales and improving fractional cover estimates, with the objective of mapping the main geological units by means of the HyMap data set. Finally, we describe and discuss our results in terms of their geological significance.

2. Field Description

2.1. Geological Setting

[11] The Sumail Massif is one of the major blocks of the Oman ophiolite (Figure 1a), a fragment of the Tethyan oceanic lithosphere formed at a spreading center during the Albo-Aptian (~95–100 Ma) [Glennie *et al.*, 1974; Tilton *et al.*, 1981; Tippit *et al.*, 1981; Ceuleneer *et al.*, 1988]. Its present setting results from intraoceanic detachment nearly contemporaneous with the accretion episode [Lanphere, 1981; Montigny *et al.*, 1988] followed by the obduction in a southwesterly direction over the Arabian continental margin, resulting in the fragmentation of the ophiolite in several blocks [Lippard *et al.*, 1986], during Maestrichtian times (~70 Ma) [Coleman, 1981; Coleman and Hopson, 1981]. Present exposure results from the relatively recent uplift of the Oman Mountains that started during the Miocene [Glennie *et al.*, 1974].

[12] The ophiolite is composed of two main lithological units: (1) a rather homogeneous mantle sequence of peridotites overlain by (2) a more heterogeneous pile of mafic-ultramafic rock types referred to as the crustal sequence [Glennie *et al.*, 1974; Hopson *et al.*, 1981; Lippard *et al.*, 1986]. The boundary between these two units corresponds to the “Moho” discontinuity of seismologists [Christensen and Smewing, 1981] (Figure 2).

[13] The mantle sequence is mainly composed of harzburgites with dunitic intervals [Braun and Kelemen, 2002], mafic veins and dykes in low abundance [Boudier and Coleman, 1981; Ceuleneer *et al.*, 1996; Python and Ceuleneer, 2003]. Two distinct magmatic series compose the crustal sequence [Juteau *et al.*, 1988]. The primary one

Figure 3. (a) Recently exposed surface of (bottom) a fresh gabbro opposed to (top) its well-developed desert varnish surface. (b) Magnesite vein network resulting from meteoric weathering affecting a well-serpentinized peridotite. (c) Preferential mechanical weathering (exfoliation) developed along joint planes yielding residual core boulders of fresh gabbro.

constitutes a classical ophiolitic stack of layered gabbro, planar-laminated gabbro, high-level isotropic gabbros, basalts in sheeted dykes and pillow lavas capped with radiolarian cherts (Figure 2). The

second one is intrusive and made of wehrlites, dunites, melanotroctolites, and rare pyroxenites cropping out as laccoliths or sills near the Moho and stocks or plugs higher up in the section.

[14] Quaternary formations, showing different depositional periods, are widespread along wadis or in terraces and alluvial fans that blanket the bed-rock (Figure 1a).

2.2. Surface Characteristics

[15] In the Sumail Massif, we encountered three broad categories of exposed surfaces that could alter its spectral analysis:

[16] 1. A desert varnish coating (Figure 3a), typically 10 to 30 μm thick on average [Dorn and Oberlander, 1982] and generally composed of a cryptocrystalline mixture of clays, ferromanganese oxides, and dust material [Potter and Rossman, 1977].

[17] 2. Secondary minerals along faults, fibrous serpentine in shear zones or along fractures in veins, and carbonate products resulting from efficient serpentine alteration (Figure 3b).

[18] 3. Exfoliation and/or abrasion of rocks generating surface constituents ranging from macroscopic boulders (3–30 cm) to microscopic mineral grains (<1 cm) which usually pile up (Figure 3c) whereas fine products accumulate down the slopes and on terraces.

3. Data Sets

3.1. HyMap Hyperspectral Image

[19] Hyperspectral images of the Sumail Massif were acquired by the HyMap airborne imaging spectrometer on 4 and 6 December 2002 [Pinet *et al.*, 2003]. The data set consists of 10×50 km flight lines oriented NE–SW. To cover the greatest possible surface, a mosaic was built using images with the lowest cloud cover and shadows. The total surveyed area covers almost 1385 km^2 with a pixel size of about 6 m (Figure 1b) acquired at 3400 m flight elevation. HyMap is a 126-channel instrument that collects data between 0.45 and 2.48 μm with a spectral resolution of 15–20 nm [see Cocks *et al.*, 1998]. Spectral coverage is nearly continuous except for the 1.40 and 1.90 μm bands that are saturated by atmospheric water vapor. HyMap data have been atmospherically corrected by HyVista on the basis of atmosphere removal radiative transfer model (ATREM) [Gao *et al.*, 1993; CIRES, 1999]

Table 1. Mean Modal Composition of Primary and Weathering Minerals Estimated for Each End-Member Lithology^a

Lithological End-Member	Primary Minerals, %						Weathering Products, %									
	Ol	Opx	Cpx	Plg	Amp	Chr	Serp	Bast	Chl	Ept	Tlc	Act	Crb	Mgt	Pre	Qtz
Harzburgite	70–90	10–30	<5			<3	20–30	<5						<5		
Dunite	95					5	75						<5 ^b	5–15		
Harzburgite with interstitial carbonates	≤10	<10				1–3	70–80						15–20 ^b	<10		
Gabbro	5–20		35–40	50–60												
Upper gabbro and dykes				50–60	40–60		≤2		≤2		≤2	≤2		<2		
Gabbro norite	<5	20–25	20–25	50–60					≤2							
Wehrlite	50–60	40–60	40–60	<10					<10	<10		<10				<5

^a Ol: olivine; Opx: orthopyroxene; Cpx: clinopyroxene; Plg: plagioclase; Amp: amphibole; Chr: chromite; Serp: serpentine minerals such as lizardite and chrysotile; Bast: bastite; Chl: chlorite; Ept: epidote; Tlc: talc; Act: actinolite; Crb: magnesite and/or dolomite; Mgt: magnetite; Pre: prehnite; Qtz: quartz.
^b Local enrichments in carbonate forming surface concretions were not taken into account but only when in veins within serpentinized peridotite samples.

and the Empirical Flat Field Optimal Reflectance Transformation (EFFORT) [Boardman, 1998].

3.2. Field Spectra

[20] To support remote sensing data, spectra were collected directly on outcrops during the overflights using a GER 3700 field spectrometer (<http://www.spectravista.com/>). Spectral resolution is 1.5 nm between 0.31 and 1.00 μm and range from 3 to 12 nm from 1.00 to 2.50 μm . The ground field of view is 5° giving a spot size of about 0.05 m² at 1 m measurement height.

[21] During a second field trip in January 2006 and using an ASD FieldSpec3[®] spectrometer (<http://www.asdi.com/>), we systematically documented all relevant rock types, including those containing unusual amounts of carbonates [see Combe *et al.*, 2006]. The ASD spectral resolution is 3 nm between 0.35 and 1.00 μm and 10 nm between 1.00 and 2.50 μm . A bare optical fiber, with a field of view of 25°, was used to analyze a surface of 0.15 m² at 1 m.

[22] For both field instruments, the diffuse reflectance of a target is determined for each wavelength by the ratio between the target's radiance and a white reference panel's radiance (Spectralon[®]) considered as a perfect Lambertian surface. Both GER and ASD data sets are also spectrally resampled to the HyMap response functions to facilitate their comparison.

4. Mineralogy and Spectral Reflectance of Ophiolite Units

[23] All rock types of the ophiolite stack and all their surface properties were directly analyzed in the field (Figure 2). We averaged site-by-site spectra of similar mineralogy and surface properties (i.e., patina, boulder size) to define end-members suitable to match pixel data. Respectively, three and four petrologic end-members were found to be the most representative geological end-members for the mantle and crustal sequences. Details about their modal fractions are presented in Table 1.

4.1. Mantle Sequence

4.1.1. Harzburgite

[24] A dark brown patina characterizes harzburgite outcrops. Serpentine minerals, common hydrothermal alterations of olivine, and to a lesser an extent of orthopyroxene, are abundant (see Table 1). In

Figure 4. Field spectra end-member of the main lithologies characterizing (a) the mantle and (b) the crustal sequence. Spectra were resampled to HyMap spectral resolution. Vertical lines indicate wavelength position of both absorption features and reflectance maximum and are discussed in the text. Atmospheric absorptions at 1.40 and 1.90 μm have been masked.

the visible near-infrared (VNIR), the mean harzburgite spectrum is dominated by a feature centered at 0.98 μm (Figure 4a) resulting from a combination of secondary goethite and limonite and primary orthopyroxene and olivine [Hunt and Salisbury, 1970; Hunt et al., 1971; Singer, 1981]. In the shortwave infrared (SWIR), it is dominated by serpentine features with typical Mg-OH absorptions at 2.13 and 2.31 μm [Hunt and Evarts, 1981; King and Clark, 1989; Clark et al., 1993; Post and Borer, 2000].

4.1.2. Dunite

[25] The exposed surface of dunite outcrops generally weathers into a very light yellowish brown to beige patina [Braun and Kelemen, 2002]. Minor olivine remains fresh in the rock (Table 1). The averaged field spectrum is quite flat in the VNIR range as often found for chrysotile, and displays typical Mg-OH features at 2.13 and 2.31 μm in the SWIR range (Figure 4a). Note here that harzburgites particularly poor in orthopyroxene tend to display similar spectra.

4.1.3. Harzburgite With Interstitial Carbonates

[26] Carbonates are weathering products created during the primary hypogenic alteration of serpentines. Secondary supergene agents provide a similar transformation process. In this case, the fluctuation of water tables slowly concentrates carbonates [Stanger et al., 1988] along fine veins (<1 mm thick) or as travertine and calcrete at the surface. Outcrops with abundant carbonate display

typical spectra of dolomite. It is therefore necessary to define a harzburgite end-member locally enriched by Mg-rich carbonates (see Table 1) [Combe et al., 2006]. Such fine carbonate veins increase the albedo of harzburgite and flatten its absorption feature at 1.00 μm but deepen the 2.31 μm one common to serpentines and Mg-rich carbonates (Figure 4a). This tends to form a symmetric peak at 2.36 μm between absorption features at 2.30 and 2.44 μm [Hunt and Salisbury, 1971] which may be used to detect samples or pixels with significant amount of dolomite-magnesite veins.

4.2. Crustal Sequence

4.2.1. Gabbro

[27] Both layered and nonlayered planar gabbros are assemblages of clinopyroxene, plagioclase, and minor olivine (Table 1) with similar crystal sizes. Hydrothermal alteration affects almost all olivine grains resulting in a mixture of serpentine, magnetite, chlorite, and talc, while plagioclase and clinopyroxene remain unaltered. The mean field spectrum collected is marked in the VNIR by a 1.03 μm broad band attributed to the ferrous ion of clinopyroxene (Figure 4b) [Clark et al., 1993]. The exact position of the second band of this mineral, generally observed near 2.40 μm , cannot be identified because of the presence of hydrothermal products with weak Fe- and Mg-OH absorptions respectively at 2.25 and 2.33 μm . The presence of fresh clinopyroxene is however detected by a maximum of reflectance at 1.65 μm [Singer,

1981; *Clark et al.*, 1993]. Despite the fact that plagioclase is abundant and fresh, it is spectrally featureless in the VNIR and SWIR range [*Hunt and Salisbury*, 1970]. As a result, reflectance spectroscopy in the 0.40–2.50 μm range may be ambiguous at discriminating among clinopyroxenite and gabbro-like spectra.

4.2.2. Upper Gabbro and Dykes

[28] Isotropic gabbros are generally rich in amphibole and locally associated with sheeted dykes striking N120–130° crosscut by N160–180° dykes. Sheeted dykes show a light gray green to green brown patina in contrast to the darker coating of isotropic gabbro. Both lithologies display a similar spectral signature due to clinopyroxene frequently altered into hornblende and interstitial Fe-Ti oxides (Table 1) in both rock types. Hydrothermal alteration results in pervasive transformation of clinopyroxene into chlorite and actinolite. Locally, strong hydrothermal alteration transforms dykes into an assemblage of epidote, albite, and quartz. The mean field spectrum displays ferric and ferrous spectral features near 0.69 and 1.02 μm . The reflectance maximum close to 1.75 μm seems to characterize the presence of amphibole [*Clark et al.*, 1993]. In the SWIR, a deep band centered near 2.35 μm corresponding to the Mg-OH vibration of both hornblende and actinolite confirms those observations. Nevertheless, both wavelength position and width of this band suggest the presence of chlorite and epidote. Both minerals are compatible with the Fe-OH feature near 2.25 μm and the Mg-OH band and shoulder located at 2.33 and 2.40 μm , respectively, masking the 2.38 μm band of amphibole. Such absorptions dominate the spectra of both rock types making these lithologies indistinguishable by VNIR-SWIR spectra and will be referred to as upper gabbro and dykes.

4.2.3. Gabbronorite

[29] Gabbronorite is characterized in the field by a brown yellow to green patina. The field spectra of this rock type displays two main absorption bands, one centered at 0.98 μm and one centered near 2.05 μm (Figure 4b). We interpret the first one as the superposition of the 0.91 μm feature of bronzite and the 1.05 μm band of diopside. As in the case of clinopyroxene, the exact position of the second band of orthopyroxene is partially masked. The presence of fresh orthopyroxene is however confirmed by a maximum of reflectance near 1.40 μm [*Singer*, 1981; *Clark et al.*, 1993]. Because fresh

plagioclases are spectrally featureless, a gabbronorite cannot be distinguished from a websterite.

4.2.4. Wehrlite

[30] A dark patina characterizes outcrops of wehrlite. This composite unit is an assemblage of olivine-clinopyroxene (i.e., a pure wehrlite) with discontinuous patches of gabbroic “impregnations” and veins (Table 1). Despite frequent weathering of plagioclase in prehnite, this mineral remains spectrally featureless, so wehrlite enriched in plagioclase and even olivine-rich gabbros cannot be distinguished from wehrlite. Olivine is serpentinized up to 30% with 5 to 15% of magnetite. Despite a degree of serpentinization similar to harzburgite and dunite, the Mg-OH 2.13 μm absorption feature is not visible in wehrlite spectra (see Figure 5a) at both sample and field scale. We hypothesize this is due to the scattering pattern of elongated magnetite grains in serpentines minimizing the interaction of light within serpentine (Figure 5b). In harzburgite, magnetite is always scarce (Figure 5c) and does not affect the spectra (see Figures 4a and Figure 5a). In dunite, when magnetite is present in similar proportions, the magnetite grains are rounded and more evenly distributed resulting in a longer path for light interaction within the serpentine (Figure 5d). Therefore, the spectrum of a rock type is not a simple linear combination of pure mineral spectra. It is the result of a complex light interaction with interlocking crystals, which is why we focused our work on rock type end-members. On the basis of powder mixtures, *Singer* [1981] demonstrated that magnetite abundance higher than 25% should mask all other absorptions features. We suggest here that spectral masking can start with 5 to 15% of magnetite in wehrlites.

5. Methods

5.1. Spectral Preprocessing

5.1.1. Airborne to Field Spectra Adjustment

[31] Dark, featureless (desert pavement), and bright (sebkha) reference targets were measured in the field, concurrently to the overflights to validate the atmospheric correction results. Differences remain between the airborne and field spectra after the atmospheric correction performed by HyVista [*Combe et al.*, 2006]. Following the *Clark et al.* [2002] procedure, we use a single reflectance to

Figure 5. (a) Sample (S) and field (F) spectra of peridotite targets within the SWIR spectral range. Dashed vertical lines highlight 2.13 and 2.31 μm absorption features related to Mg-OH. Scanning Electron Microscope (SEM) images of (b) wehrlite, (c) harzburgite, and (d) dunite of the areas corresponding to sample spectra. Ol: olivine; Serp: serpentine; Mgt: magnetite.

compensate this effect by adjusting the image against the bright surface spectrum (see Figure 6) following equation (1):

$$\rho_{calib} = \rho_{img} \cdot \frac{\rho_{bf}}{\rho_{bi}} \quad (1)$$

where ρ_{img} is the reflectance spectrum of a given pixel, ρ_{bf} the mean spectrum of the bright surface measured on the field and ρ_{bi} the mean spectrum of the same target acquired by HyMap (Figure 6).

5.1.2. Surface Scattering and Shadowing Effects Removal

[32] Photometric variations in a high-relief mountainous area such as in the Sumail Massif may induce albedo and spectral shape changes. As an example, a sunlit gabbro surface displays a slightly more convex spectrum of reflectance with higher mean intensity than a partially shadowed surface spectrum of the same lithology (Figure 7a). The same kind of differences exists between airborne image and field spectra (Figure 7b) or between targets acquired at different scales [Friedl *et al.*, 1995; Chen, 1999].

[33] Other photometric variations also arise because of grain size and distribution variation between lithologies having similar mineralogical compositions [e.g., Cord *et al.*, 2005]. In the

Figure 6. Reflectance spectra of a sabkha homogeneous surface acquired in the field (gray) and observed by the HyMap instrument (black) used to adjust the airborne data set (for more details, see Combe *et al.* [2006]).

Figure 7. (a) Sunlit and shadowed surface gabbro spectra acquired by HyMap under natural sunlight. (b) Homogeneous gabbro surface spectra acquired by both ASD and HyMap spectrometers at different scales with the same light source. (c and d) Trend curves resulting from a Gaussian low-pass-filter ($\tau = 1.40 \mu\text{m}$) of spectra shown on Figures 7a and 7b. (e and f) Normalized spectra for the two previous cases showing a good match between all samples confirmed by a quite flat residual spectra.

0.40–2.50 μm wavelength range, fine grain sizes tend to increase the albedo of a lithology while large grain sizes induce deeper absorption features (e.g., pyroxenes in the work of *Singer* [1981] and *Clark et al.* [1993]).

[34] Effects on the albedo and the overall shape of the spectra can be isolated from absorption features by considering only the upward limit of the overall reflectance curve which is called continuum by

Clark and Roush [1984] and *Mustard and Sunshine* [1999]. An approximation of the continuum is usually proposed by means of a tangent line linking the reflectance maxima and bridging over all absorption features. But, as stated by *Roberts and Herold* [2004], it can be problematic at the wavelength boundaries used to define that continuum. Absorption features near 0.40 μm and 2.50 μm are also missed. Accordingly, on the basis of this approach [*Clark and Roush*, 1984], a standard

Figure 8. Discrete wavelength intervals illustrating weak shape differences (a) in the SWIR and (b) in the VNIR of both mantle and crustal sequences end-members. Bands used by *Combe et al.* [2006] are indicated by gray arrows. Bands used in this study to perform the partition index I_p are indicated by black arrows.

normalization of each spectrum to its continuum shape cannot be achieved in the present study. Moreover, ideally we wish to remove only the optical contribution arising from the surface state and not the contribution due to rock-type internal grain sizes. For this purpose, *Combe et al.* [2005, 2006] proposed to subtract a continuum fraction based on a straight-line function in logarithm reflectance space and in wave number as established by the MGM procedure [*Sunshine et al.*, 1990]. This procedure requires fixing the tie points for the modeled continuum as input parameters. However, this constraint can cause major problems when applied to images where spectral shapes vary from one pixel to another because of changes in composition. In particular, we have shown here that the observed reflectance maxima can shift depending on rock type. Therefore, based on the differences noted between spectra (e.g., Figures 7a and 7b), we developed a new method to estimate a

continuum better suited to our case, based on the trend curve of the spectrum and given by a basic Gaussian low-pass filter. The Gaussian coefficients of such a filter are given by equation (2).

$$A(\lambda) = 0.5 \left(\frac{\lambda - \lambda_0}{\tau} \right)^2 \quad (2)$$

where A is the amplitude of the Gaussian curve, λ_0 is its band center, τ is its full width at half maximum (FWHM), and λ is the wavelength. Figures 7c and 7d show the trend curve spectra of Figures 7a and 7b, respectively. The division of each spectrum (ρ_s) by its trend curve (ρ_t) for each wavelength gives a normalized reflectance (ρ_n) as follows (equation (3)) facilitating the comparison between spectra.

$$\rho_n = \frac{\rho_s}{\rho_t} \quad (3)$$

Figures 7e and 7f illustrate two applications showing that this procedure allows a perfect match

Figure 9. Composite image of three mantle sequence end-members fraction cover maps, where H is harzburgite, D is dunite, and H⁺C is harzburgite with interstitial carbonate (see spectra Figure 4a). (a) Fraction covers calculation by comparison between each pixel spectrum to the synthetic spectral library after Gaussian low-pass trend normalization. (b) Standard linear spectral unmixing of field rock type end-member spectra extracted from each pixel spectrum of the image.

between sample spectra acquired at different scales without altering their signature, but at the expense of the albedo information which is lost.

5.2. Spectral Partition of the Ophiolite Sequence

[35] To facilitate image processing, we found it useful to partition the image data into two independent subsets. *Combe et al.* [2006] defined a simple index built on the characteristic M shape (see gray arrows in Figure 8a) of Mg-OH bands combination to determine whether a pixel belongs to the mantle or to the crustal sequence. However,

because of the noise level of some dark parts of the image we rather used an index, based on the curvature of the triplet 2.16–2.20–2.31 μm which varies with increasing Mg-OH content from harzburgite, (straight line), to dunite, (slightly convex curve), as shown in Figure 8a (black arrows), while crustal lithologies display concave curves in Figure 8b. To avoid ambiguities in the region of the Moho between dunite and wehrlite, we combine this “Mg-OH index” with one based on the curvature of the triplet 0.96–1.07–1.20 μm related to the presence of iron (see black arrows in Figure 8b). The resulting “partition index” is the following (equation (4)):

$$I_p = \sqrt{\frac{1}{4} \frac{(B_{1.20} + B_{0.96})(B_{2.16} + B_{2.31})}{B_{1.07} \cdot B_{2.20}}} \quad (4)$$

with the crustal sequence presenting higher index values than the mantle sequence.

5.3. Rock Type Fractional Cover Mapping

[36] Basically, a geological map presents continuous areas covered by a main rock type. When rock type imbrications exist at scales less than the pixel size, one should consider defining intermediate mixtures of rock types for establishing maps. However, different levels of mixture exist and should be addressed prior to any mapping.

[37] For instance a simple mixture occurs when a pixel is covered by two different lithologies. The spectrum of a pixel is in this case a simple linear combination of each lithology spectrum, proportional to its fractional cover. However, intimate mixtures also occur when photons interact with more than one mineral at a grain scale and are generally nonlinear [*Nash and Conel*, 1974; *Singer*, 1981]. Therefore, a rock spectrum is not a linear combination of single mineral spectra. As a result, the occurrence of two rock types in a pixel can display a spectrum which may look like a third rock type, which is the case of most end-members identified in both mantle and crustal sequences. Consequently, any attempt at unmixing linearly [e.g., *Adams et al.*, 1986; *Boardman*, 1989] such complex end-members may lead to incorrect determinations. To interpret these complex spectra it is necessary to use radiative transfer theory such as the one developed by *Hapke* [1981, 1993] which takes into account the exact imbrications of minerals and their surface optical properties [e.g., *Cord et al.*, 2005]. However, we choose here a simpler strategy developed for biomass quantification on

Figure 10. Diagrams illustrating the classification method applied to develop the geological map. It is based on the combination of the fraction cover given by (a) a ternary or (b) a quadruple lithological association. End-members are located at the summits of triangles while associations are on the edges. When the three or four lithologies display similar proportions, none is dominant and corresponding pixels are defined as undifferentiated lithologies (U.L.).

mudflats [Mélédér *et al.*, 2003; Combe *et al.*, 2005]. We build a spectral library containing all possible surface combinations between rock types and constrain the analysis to the detection of the exact match between pixel and each combination of the library. For example, this procedure allows to better estimate harzburgite in the mantle sequence (Figure 9a) whereas spectral mixture analysis, based on a simple mixture model, results in an overestimation of the carbonate content (Figure 9b).

[38] One can build an N end-member spectral library with the following equation (5).

$$\rho_{mix}(\lambda) = \sum_{i=0}^N x_i \cdot \rho_i(\lambda) \text{ with } \sum_{i=0}^N x_i = 1 \quad (5)$$

where ρ_{mix} is the reflectance of the mixed spectrum, ρ_i is the reflectance of the surface end-member i , and x_i is the fraction of each constituent. The increment used is 0.1 thus it creates a library of 66 spectra in the case of three end-members and 286 spectra with four end-members.

[39] As shown on Figure 7, image and field spectra match when divided by their Gaussian low-pass trend. Therefore, we divide each spectrum of the library by its Gaussian low-pass trend. We compare the spectrum of each pixel to each spectra of the spectral library using a spectral distance (SD) which consists in calculating the absolute difference in length between vector v and w as given by equation (6) [Bakker and Schmidt, 2002].

$$SD(\vec{v}, \vec{w}) = \|\vec{v} - \vec{w}\| \quad (6)$$

Each pixel spectrum distance to a library spectrum is given by equation (7).

$$SD(\vec{v}, \vec{w}) = \sqrt{\sum_{\lambda} [\rho_v(\lambda) - \rho_w(\lambda)]^2} \quad (7)$$

where λ is the wavelength of the n bands, ρ_v is the reflectance spectrum of the pixel, and ρ_w is that of the library spectrum. The SD calculation is constrained to the 0.54–2.40 μm spectral interval in order to remove the noise present at both ends of the airborne spectra. For subtle variations between library and pixel spectra SD varies from 1 to 0, whereas unrelated spectra produce unrealistically large SD variations. In practice we rather use the inverse distance (equation (8)) that reveals subtle variations on visual displays.

$$I_d = \frac{1}{SD} \quad (8)$$

I_d tends toward zero for spectra that are completely different from each other and becomes sensitive to small differences when spectra are similar (i.e., values greater than 1). To avoid I_d becoming infinite, we exclude all distances (SD) smaller than 0.0001. The greatest I_d value designates the best match between the spectrum of a pixel and one from the library. Pixels having an I_d smaller than 1.5 are removed to avoid unmodelled spectra or unknown lithologies. This I_d threshold corresponds to the minimum value computed for the seven ground control points and their corresponding airborne spectra.

Figure 11. (a) Composite image of HyMap (red indicates 2.30 μm ; green indicates 2.16 μm ; blue indicates 2.02 μm). (b) The same HyMap composite image with normalized reflectance. The white dashed bold line indicates the boundary, which is the petrologic Moho. Arrows identify interesting features and are discussed in the text.

5.4. Geological Mapping by End-Member Fraction

[40] In this section, we combine the end-member fraction cover maps into one synthetic map. A pixel is assigned to a petrologic rock-type when its corresponding end-member contributes to more than 60% to the spectrum of the pixel on the fraction cover image, following a rule of classification shown on Figure 10a for three end-members and Figure 10b for four end-members. When two end-members make up to 80% of a pixel, typically 40% each, it is set to be a characteristic association of both petrologic rock types. Pixels where no lithology reaches 40% are considered undefined and set as unidentified lithologies. This geological classification scheme contains seven end-members

for the mantle sequence and 11 end-members for the crustal sequence.

6. Results

6.1. Lithological Maps

[41] The Gaussian low-pass trend normalization allows us to quickly visualize the mantle and crustal sequences in a simple color composite image using three SWIR channels, at 2.30 (red), 2.16 (green), and 2.02 μm (blue) as shown in Figure 11. These three channels are chosen as they are characteristic of serpentine and clay minerals but relatively insensitive to spectral broadening in the range 1.80 to 2.00 μm due to absorption of variable amounts of atmospheric water vapor.

Figure 12. Normalized HyMap spectra of the main lithological end-members. The gray vertical lines mark the three bands used to visualize the ophiolite sequence in Figure 11.

The mantle sequence appears in “green cyan” (Figure 11b), as a result of the drop of reflectance near $2.30 \mu\text{m}$ characterizing (serpentinized) harzburgites and dunites (Figure 12), while the crust (mostly free of serpentine) appears “red magenta” (Figure 11b). Field and petrographic studies confirmed that “magenta” pixels are mainly olivine gabbros (Figure 11b, arrow A), “purple” pixels are gabbros richer in pyroxene (B), “violet” pixels are wehrlites (C), and “white pink” pixels are intrusive diabbases dykes (D). To the south of Muqzah, a crustal block appears more “reddish” (E), suggesting the occurrence of orthopyroxene, inducing a positive spectral slope from 2.00 to $2.50 \mu\text{m}$ (Figure 12). Sedimentary formations of the Jabal Al Akhdar appear in “white pink” in the southwestern edge of the image (F) and have been removed in Figures 13–18.

[42] Note that such a composition uses three channels included in a single Landsat TM channel (TM7). The three closest ASTER channels (Band 6: $2.18\text{--}2.22 \mu\text{m}$; Band 7: $2.23\text{--}2.28 \mu\text{m}$; Band 8: $2.29\text{--}2.36 \mu\text{m}$) are shifted in wavelength and broader than the ones we used here thus displaying less contrasted and noisier results due to the limited number of channels available to calculate an equivalent robust Gaussian low-pass trend.

[43] Products of erosion sometimes undergo significant transport during which they are abraded and often show fresher minerals than at their source. Consequently, Quaternary formations result from the accumulation of mantle and crustal materials deposited over other rock types. Some wadi beds are composed of float of crustal lithology (Figure 11b, arrow G) while others are composed of mantle lithologies (H). Thus, it is impossible to

map them without geological interpretation. Masking those Quaternary deposits, we are able to outline the Moho limit (Figure 11b) at the base of the transition zone.

6.2. Rock Type Fractional Cover Maps

6.2.1. Spectral Partition of the Ophiolite Sequence

[44] The spectral resolution and the number of bands of the HyMap data permit us to highlight a specific mineralogy using a specific index. This is

Figure 13. (a) I_p greyscale image showing the distribution and extent of the mantle (dark) and the crustal (bright) sequences. (b) Histogram illustrating the distribution of I_p according to the number of pixels. The mantle and the crustal sequence are highlighted by a bimodal distribution. The dashed vertical line emphasizes the threshold value of 0.97 used to split the image into two ophiolite units (i.e., mantle and crust).

Figure 14. Fraction cover maps of mantle end-members: (a) harzburgite, (b) dunite, and (c) harzburgite with interstitial carbonates (see code bar for scale). The Moho boundary is indicated in normal dashed line. Arrows identify interesting features and are discussed in the text.

achieved here with the partition index map I_p (Figure 13a) showing a bimodal distribution (Figure 13b). It helps to divide the data set into mantle and crustal sequences based on Mg-OH and Fe modulations on the spectra curvature. This partition allows us to build dedicated mixing models for both the mantle and the crustal sequences and to reduce confusion between rock types. As an example, a wehrlite spectrum is close to a dunite spectrum except for the narrow bands used for the partition. Without such a procedure, a number of ambiguities would prevent mapping of these two rock types.

6.2.2. Three End-Member Model of the Mantle Sequence

[45] Harzburgite is the most abundant lithology of the mantle sequence, in the NNE as well as in the SSW of the Massif (Figure 14, arrows A), as expected from previous geologic mapping [Amri,

1995]. Both areas are dominated by harzburgite, with fractions decreasing from 100% to 70% at their periphery. On the other hand, in the Maqсад area, the proportion of harzburgite relative to dunite decreases (B), with dunite proportion exceeding 60% (C). Dunite is also found in the eastern side (D) lying beneath the petrologic Moho. Moreover, fractions of each rock type ranging from 40 to 60%, highlighting two parallel structures oriented NW–SE (E). Finally, harzburgite with interstitial carbonates is generally scattered in other units (F) but its relative abundance reaches 80% in restricted areas near Muqbariyah (G) or along parallel structures striking N30–40° (H).

6.2.3. Four End-Member Model of the Crustal Sequence

[46] The map of spectral distance between image and gabbro-norite ground spectra (Figure 15) shows that gabbro-norite appears only in a block south of

Figure 15. I_d map of the gabbronorite end-member showing its presence in the southern part of the crust.

Muqzah. To avoid unnecessary calculations, we compare this block to a four end-member library including gabbronorite, and we compare the rest of the crustal section to a three end-member library containing the same three end-members considered above: gabbro, wehrlite, and upper gabbro and dykes.

[47] The gabbro unit (Figure 16a) appears in spots with cover fractions greater than 70%, near Nidab (Figure 16, arrow A), Mahram (B), northeast of Sayma (C), and Hamitha (D), and in the northwest of the southern blocks (E). The upper gabbro and dykes unit displays proportions ranging from 40 to

70% along lineaments crosscutting the whole crust in a NE–SW direction in the north (F) and in a NNW–SSE direction in the south (G). The wehrlite unit (Figure 16c) is observed in large corridors oriented NW–SE (Figure 16c, arrow H), as stocks (I), or as long and narrow structures striking NE–SW (J). Areas having proportions of wehrlite ranging from 20 to 40% are more common and often confused with gabbro outcrops (K). It seems that multiple narrow injections of wehrlite in gabbro tend to be mapped as the wehrlite end-member. In the southern block, gabbro (E) and wehrlite (L) alternate with gabbronorite (Figure 17, arrow M) cropping out in narrow parallel bands ranging from 10 to 100 m in thickness.

6.3. Geological Maps

[48] Maps of rock type cover have been merged into one geological map following the procedure presented in § 5.4. Harzburgite is the main constituent of the mantle sequence. The Muqbariyah Shear Zone [Amri *et al.*, 1996] would appear to be characterized by abundant dunite. However, these dunite-like spectra may be caused by the enhanced serpentinization of orthopyroxene in the shear zone. The harzburgite enriched in carbonates, detected by Combe *et al.* [2006], mainly appears scattered in serpentinized areas and the highest concentrations are found along wadis revealing old water table fluctuations confirmed by field observations of calcrete. Clinopyroxene-rich intrusions aligned along N120° directions are known in the northern part of the Sumail Massif [Amri, 1995]. The present survey allowed us to determine more accurately the boundaries of some of these intrusions, among the richest in clinopyroxene (Figure 18b), and agree with the results found with a different approach by Pinet *et al.* [2006b].

[49] The crustal sequence is more complex because of amphibole-clinopyroxene-orthopyroxene-olivine mixtures in numerous rock types. For this reason, upper gabbros and dykes are grouped in one end-member characterized by amphibole abundance. As mentioned by Combe *et al.* [2006] and Amri [1995], gabbro and olivine gabbro appear to be the dominant lithologies in the crustal section (Figure 18). However, the widespread detection of wehrlite-bearing pixels (Figure 17c) is consistent with field observations of abundant interlayered wehrlite in gabbros. Such a wehrlite-rich horizon crops out along a NW–SE corridor (Figure 18c), interpreted by Amri [1995] as the paleosspreading ridge axis.

Figure 16. Fraction covers maps of crustal end-members: (a) gabbro, (b) upper gabbro and dykes, and (c) wehrlite (see code bar for scale). The bold dash-dotted frame indicates the area treated with four end-members, instead of three end-members. Arrows identify interesting features and are discussed in the text.

[50] A gabbro-norite unit characterizes the crustal sequence SW of the Muqbariyah shear zone which is a major structural feature separating two segments of the Oman ophiolite. In the mantle sequence, *Python and Ceuleneer* [2003] mapped the distribution of the various types of igneous dykes interpreted as channels for the melts feeding the overlying magma chambers, where the crustal cumulates eventually crystallized. In the Maqсад area, to the NE of the Muqbariyah shear zone, these channels are made of troctolites (olivine + plagioclase) and of olivine gabbros; these are the crystallization products of MORB-like melts (olivine tholeiites) produced within the Maqсад diapir. In the mantle section SW of Muqbariyah, mantle dykes are of websteritic and gabbro-noritic composition. These are the crystallization products of melts richer in silica than MORB, typically of andesitic composition. The abundance of orthopyroxene in the crustal section SW of Muqbariyah,

suggested by the present spectroscopic study, and its absence from the crustal section related to the Maqсад diapir (present study and *Amri* [1995]), are consistent with the distribution of the various types of mantle dykes [*Python and Ceuleneer*, 2003]. This result confirms that the mantle and the crust, in a given area of the Oman ophiolite, have experienced a common tectonic and petrogenetic evolution.

[51] Quaternary deposits cannot be simply distinguished from their parent rock based on spectral signatures. Such deposits are generally a mixture of a wide range of lithologies, from gabbro to harzburgite. Consequently, the spectra of such areas may correspond to unexpected mixture of rock types. This fact explains for instance why we encountered a mixture of wehrlite and dykes spectra in the alluvial fan near Muqbariyah (Figure 18a) or wehrlites in the M-shape wadi near Mahram

Figure 17. Fraction cover map of the gabbronorite end-member (see code bar for scale). This fourth end-member refers to framed details shown in Figures 16a, 16b, and 16c. The bold dash-dotted line indicates the boundary of the area processed with three end-members. The normal dashed line underlines the petrologic Moho. Arrows identify features which are discussed in the text.

(Figure 18d). The distinction between igneous and sedimentary outcrops must be made by the geologist on a geomorphological basis.

7. Conclusion

[52] A new mapping strategy was developed and applied to a HyMap survey of the Sumail Massif. It is mainly based on the direct comparison of the spectrum of each pixel with reference spectra. These reference spectra are taken from synthetic spectral libraries, built by the combination of selected rock type end-member spectra acquired in the field.

[53] In most places, outcrop exfoliation leads to the exposure of minerals sufficiently fresh to give spectral responses diagnostic of primary igneous rock types. Secondary minerals inherited from hydrothermal alteration may, however, modify some absorption features to an extent that depends on the (primary) igneous minerals assemblage. For example, serpentine is responsible for deep absorption features in harzburgite and dunite but in wehrlite it is as a consequence of the greater abundance of magnetite crystals in the wehrlite than in both harzburgite and dunite. One must bear in mind the spectral signature of a given rock type

is the result of a nonlinear combination of spectral signatures of imbricated minerals. Furthermore, some minerals such as plagioclase are not detectable in the wavelength range used, which makes the distinction between gabbro and clinopyroxenite, or between gabbronorite and websterite, impossible. A careful selection has thus been made to define the rock type end-members that are effectively detectable by spectroscopy. Despite this selection, standard inverse methods, such as linear unmixing, have not been able to retrieve the real proportion of each rock type in the field. This is due to the fact that intermediate combinations of two rock types may be spectrally close to a third rock type. We avoided this difficulty through the use of a dedicated reference spectral library including all the possible combinations of rock types present in the field. This procedure was indeed quite successful in detecting the presence of carbonate in harzburgite while the inverse method tended to overestimate the carbonate amount.

[54] On the basis of the detection of a narrow absorption feature characteristic of the presence of serpentine, we were able to split the data into two separate subsets, dealing with the mantle and crustal sequences, respectively. On the basis of a simple measure of the degree of similarity between spectra divided by their Gaussian low-pass trend, the use of appropriate end-member combinations for each section enhances the ability to provide geologically meaningful maps.

[55] On the basis of these results, maps of relative rock type abundance have been constructed. In the mantle sequence, we distinguish a lower horizon composed of harzburgite from an upper horizon composed of dunite and depleted harzburgite. In the crustal sequence, this work reveals the presence of orthopyroxene in a block south of a major shear zone (Muqbariah) while this mineral is absent from the crustal section north of Muqbariah, a spectral observation consistent with independent structural and petrologic observations. Our survey mapped the distribution of wehrlite bodies and amphibole enriched gabbro and dykes efficiently and objectively.

[56] Though plagioclase cannot be unambiguously detected by reflectance spectroscopy, the comparison between the proposed geological map and those established earlier by traditional mapping methods indicates that hyperspectral remote sensing is an adequate tool for geological mapping requirements: as an example, the determination of igneous rock type boundaries is clearly improved.

Figure 18. Geological map of the studied part of the Sumail Massif. (a to e) White frames highlight interesting areas that are discussed in the text. Abbreviations of the defined lithostratigraphic units: H, harzburgite; HD, harzburgite and dunite; HC, harzburgite with secondary carbonates; HiC, harzburgite with interstitial carbonates; DHC, dunite and harzburgite with interstitial carbonates; D, dunite; UM3, undifferentiated mantle rocks defined by three end-members; G, gabbro; GW, interlayered gabbro and wehrlite; W, wehrlite; GnW, interlayered gabbro and wehrlite; Gn, gabbro; GGn, interlayered gabbro and gabbro; GDy, gabbro with intrusive dykes; WDy, wehrlite with intrusive dykes; GnDy, gabbro with intrusive dykes; UGD, upper gabbro and dykes; UC3, undifferentiated crustal rocks defined by three end-members; UC4, undifferentiated crustal rocks defined by four end-members; U, unmapped as I_d is smaller than 1.5 or unknown.

Its sensitivity to narrow features as well as slight variations of the shape of spectra leads to a good discrimination of subtle variations in mineralogy in a unit that would have been considered homogeneous with a traditional approach. The strategy presented here mainly relies on geological consid-

erations, close to those used when constructing a geological map, and we believe that as hyperspectral data become available, such a mapping procedure should be used wherever possible. We emphasize the need for in situ measurements and a preliminary knowledge of the geological context.

We also point out the fact that this method is sensitive to image quality. We have shown, however, that under degraded conditions, the amount of unmappable areas (black in Figure 18) could be reduced, and that a moderate amount of shadow-darkened area is not too severe a limitation for the production of accurate lithological maps.

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