Condensed matter clouds treated in grand canonical ensemble

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All kinds of clouds are condensed matter and their various thermodynamic functions can be derived successfully by statistical methods. First, we will discuss cloud behaviour in regulating the heat transfer in the M (meso-), S (strato-) and T (troposphere) and its effect on the surface temperature of the Earth. Next, a simple method is evolved to calculate the Earth's surface temperature and finally, clouds are considered as condensed matter treated in the grand canonical ensemble.

Keywords: Clouds, condensed matter, Earth's surface temperature, grand canonical ensemble.

The Earth is my mother and I am her son.

Upanishad

THE study of clouds, where they occur, and their characteristics, play a key role in understanding climate change. Low, thick clouds primarily reflect solar radiation and cool the surface of the Earth. High, thin clouds primarily transmit incoming solar radiation; at the same time they trap some of the outgoing infrared radiation emitted by the Earth and radiate it back downwards, thereby warming the surface of the Earth. Whether a given cloud will heat or cool the surface depends on several factors, including the altitude of the cloud, its size and make-up of the particles that form the clouds. The balance between the cooling and warming actions is very close although, overall, averaging the effects of all the clouds around the globe, cooling predominates[1,2].

The Earth's climate system constantly adjusts in a way that tends towards maintaining a balance between the energy that reaches the Earth and the energy reflected back to space. The composition of the Earth system that is important to the radiation budget includes the planet's surface, atmosphere and clouds. Over the whole surface of the Earth, about 30% of incoming solar energy is reflected back to space. Because a cloud usually has a higher albedo than the surface beneath it, the clouds reflect more shortwave radiation back to space than the surface would in the absence of the clouds, thus leaving less solar energy available to heat the surface and atmosphere. Hence this 'cloud albedo forcing' taken by itself tends to cause cooling or 'negative emission forcing' of the Earth's climate. When a cloud is introduced into a previously clear sky, the cold cloud top will reduce the long warmth mission to space and (disregarding the cloud albedo forcing for the moment) energy will be trapped beneath the cloud top. This trapped energy will increase the temperature of the Earth's surface and atmosphere until the longwave emission to space once again balances the incoming absorbed shortwave radiation. This process is called 'cloud green house forcing' and taken by itself tends to cause a heating or 'positive forcing' of the Earth's climate[3].

Usually the higher a cloud is in the atmosphere, the colder is its upper surface and greater is its cloud greenhouse forcing.

If the Earth had no atmosphere, a surface temperature far below freezing would produce enough emitted radiation to balance the absorbed solar energy. But the atmosphere warms the planet and makes the Earth more inhabitable. In addition to the warming effect of clear air, clouds in the atmosphere help to moderate the Earth's temperature. The balance of the opposing cloud albedo and cloud greenhouse forcing determines whether a certain cloud type will add to the air's natural warming of the Earth's surface or produce a cooling effect. The high, thin, cirrus clouds tend to enhance the heating effect and the low, thick, stratocumulus clouds have the opposite effect, while deep convecting clouds are neutral. The overall effect of the clouds together is that the Earth's surface is cooler than it would be if the atmosphere had no clouds.

The high, thin, cirrus clouds in the Earth's atmosphere act in a way similar to clear air because they are highly transparent to shortwave radiation (their cloud albedo forcing is small), but they readily absorb the outgoing longwave radiations.

In contrast to both the cloud categories are deep convective clouds typified by cumulonimbus clouds. A cumulonimbus cloud can be many kilometres thick, with a base near the Earth's surface and there is essentially no excess radiation. In consequent of the cloud greenhouse forcing and albedo forcing the overall effect of cumulonimbus clouds is neutral.

The Earth's surface albedo A_E exhibits strong correlation with the albedo of clouds A_C. Their direct relationship can be derived in an intuitive manner.

Let $F_u^{\downarrow}$ be the downward flux of shortwave radiation (direct solar and diffuse) at the top of the upper cloud boundary, $R_u^{\uparrow}$ be the reflected flux from the upper surface of the cloud, $F_L^{\downarrow}$ the downward flux of shortwave radiation at the lower boundary of the cloud and $F_L^{\uparrow}$ the upcoming flux of shortwave radiation (diffuse atmospheric or reflected by the Earth's surface) at the lower cloud boundary level. Now for the cloud albedo A_C, the relative values of the transmission coefficient P and absorption coefficient $(1 – P)$, we have the following relations (Figure 1)

$$A_C = \frac{R_u^{\uparrow}}{F_u^{\downarrow}},\; A_E = \frac{F_L^{\uparrow}}{F_L^{\downarrow}},\; P = \frac{F_L^{\downarrow}}{F_u^{\downarrow}},$$

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$$(1-P)=\frac{F_u^{\downarrow}+F_L^{\uparrow}-R_u^{\uparrow}-F_L^{\downarrow}}{F_u^{\downarrow}}=1+A_E\cdot P-A_C-P\,,$$

$$\therefore P=\frac{A_C}{A_E}. \tag{1}$$

The transmission coefficient P has also been defined as

$$P=\frac{G}{\mathrm{ETR}}=a+b\left(\frac{n}{N}\right),$$

where G is the total solar radiation at the ground and ETR is the extra terrestrial radiation at the top of the atmosphere. a and b are regression constants. The cloudiness can be defined as[4]

$$C=1-\left(\frac{n}{N}\right), \tag{2}$$

where n is the daily duration of sunshine, N the maximum possible duration of sunshine, a and b are established empirically for each location under Indian climatic conditions, $0.14 \le a \le 0.54$; $0.18 \le b \le 0.73$ and $0.60 \le (a + b) \le 0.85$. The magnitude of a depends upon the type and thickness of the prevailing clouds and that of b upon the transmission characteristics. $(a + b)$ is the overall mean transmission factor for daily G under clear sky conditions. Lower values of a are invariably associated with higher values of b and vice versa. Figure 2 shows relation between $P = A_C/A_E$ versus C. The values of P and C are calculated from eqs (1) and (2).

It is assumed that the infrared absorption by the cloud is equal to its emissivity. The downward radiations from the cloud add to the surface heating, thus contributing to the greenhouse effect. Absorbed and emitted or reflected radiations at each level (upper and lower boundary of the cloud layer and the Earth's surface) are equated to yield the surface temperature T_s given by[5]

$$\sigma T_S^4=\frac{S(1-A_C)(2-a_C)}{(2-\varepsilon)}, \tag{3}$$

where a_C is the absorption coefficient of the clouds equal to $1 - P$.

Equation (3) can be written as

$$\sigma T_S^4=\frac{S(1-A_C)(1+P)}{(2-\varepsilon)}. \tag{4}$$

Using eqs (1) and (2) in eq. (3), we get

$$\sigma T_S^4=S\,(1-A_C)\{(1+(a+b)-bC)\}/(2-\varepsilon). \tag{5}$$

From eq. (5) one can calculate the Earth's surface temperature in terms of cloud albedo A_C, cloudiness C and emissivity (Table 1)[5].

The cloud albedo $A_C = 0.80$ for $C = 0$ due to Rayleigh scattering in the atmosphere. S is the quarter of the solar flux = 343 $\mathrm{Wm^{-2}}$. $\varepsilon = 0.4$ (for $C = 0$) and 0.9 (for $C = 1$) and the Stefan–Boltzmann constant $\sigma = 5.67 \times 10^{-8}\ \mathrm{wm^{-2}\ k^{-4}}$.

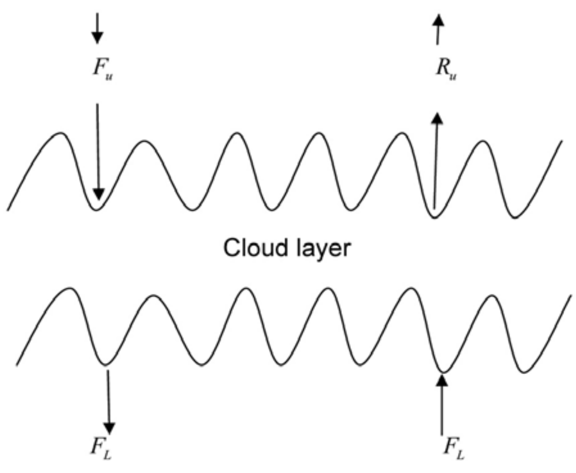

Figure 1. Cloud model.

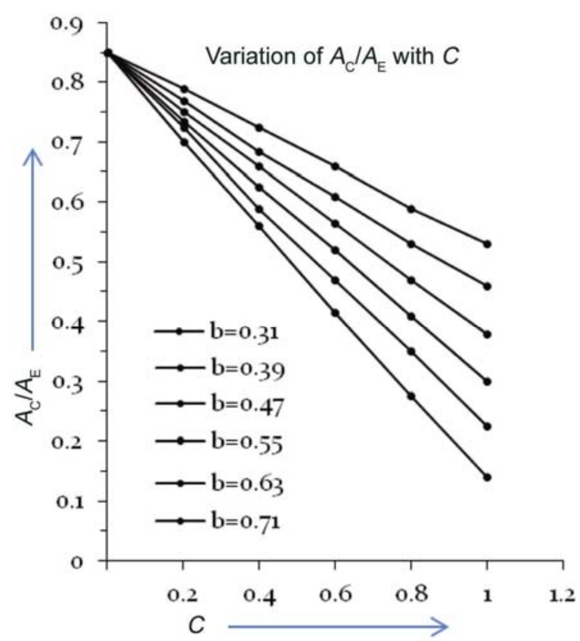

Figure 2. Variation of A_C/A_E with cloudiness C.

Table 1. Albedo of clouds

Cloud level	Albedo
High (Ci, Cs, Cc)	0.25
Middle (Ac, As)	0.60
Low (Ns, St, Sc, Cu and Cb)	0.70

Table 2. Calculation of the Earth's surface temperature and variation in cloudiness

Cloudiness *C*	*a* + *b*	*b*	Cloud albedo A_C	Emissivity	Surface temperature *T* (K)
0	0.6	–	0.08	0.40	273.13
0	0.85	–	0.08	0.40	283.2
1	(*a*)0.14	–	0.25	0.90	261.9
1	0.14	–	0.6	0.90	223.8
1	0.54	–	0.7	0.90	208.3
0.5	(*a* + *b*)0.76	0.18	0.25	0.90	288.16 (average temperature)
0	0.60	–	0.08	0.90	299.95
0	0.85	–	0.08	0.90	311.04

The different values of T_S calculated after carrying out sensitivity tests using eq. (5) are given in Table 2.

In Table 2, the surface temperature T_S has been calculated in some extreme cases of cloudiness *C*, the regression constants *a* and *b*, cloud albedo A_C and infrared emissivity ε. The mean temperature T_S for partial cloudiness ($C = 0.5$) and nearly mean value of ($a + b$) has been calculated as 288.16 K, which is close to the observed value. The greenhouse effect is seen for $C = 0$ and $a + b = 0.60$ and 0.85. The maximum temperature permissible for maximum $\varepsilon = 0.90$ is 311.04 K, provided the solar constant *S* does not vary. The upper-level clouds cause warming and lower level clouds cause cooling as observed.

Atmospheric thermodynamics is the study of heat to work transformations (and the reverse) in the Earth's atmospheric system in relation to weather or climate. Following the fundamental laws of classical thermodynamics, atmospheric thermodynamics studies such phenomena as properties of moist air formation of clouds, atmospheric convection, boundary layer, meteorological and vertical stabilities in the atmosphere. Atmospheric thermodynamic diagrams are used as tools in the forecasting of storms. Atmospheric thermodynamics forms a basis for cloud microphysics and convection parameterization in numerical weather models and is used in many climate considerations, including convective-equilibrium climate models[6].

Atmospheric thermodynamics focuses on water and its transformation. Areas of study include the law of energy conservation, the ideal gas laws, specific heat capacities, adiabatic processes (in which entropy is conserved) and moist adiabatic processes. Most of the tropospheric gases are treated as ideal gases and water vapour is considered as one of the most important trace components of air.

Advanced topics are phase transitions of water, homogenous and inhomogeneous nucleation, effect of dissolved substances on cloud condensation, role of supersaturation on formation of ice-crystal and cloud droplets. Considerations of moist air and cloud theories typically involve various temperatures, such as equivalent potential temperature, wet-bulb and virtual temperature. Connected areas are energy, momentum and mass transfer, turbulence interaction between air particles in clouds, convection, dynamics of tropical cyclones and large-scale dynamics of the atmosphere[7].

The major role of atmospheric thermodynamics is in terms of adiabatic forces acting on air parcels, including in primitive equations of air motion either as grid-resolved or sub-grid parameterizations. These equations form a basis for the numerical weather and climate predictions.

The dimensions of cloud water droplets are comparable with the wavelength of the thermal radiation. In this case, therefore, it is important to consider scattering. The exact solution of the problem of radiative heat transfer in clouds can be obtained only by using the exact transfer equation.

It has been pointed out that in the considered case the absorption coefficients, like the scattering coefficients, can be correctly determined only by applying the diffraction theory of the radiative properties of cloud droplets. It appears, for example, that by virtue of the diffraction effects, the radiation by a droplet in the region of certain wavelengths exceeds the black radiation at the temperature of the droplet. The calculation of scattering and absorption coefficients makes possible an adequate solution of the problem of radiative heat transfer in lower level clouds containing water droplets of definite radius.

The main conclusion concerning thick stratus clouds is that a cloud is 'active' with respect to thermal radiation only near its edges, that is, in the boundary layers. The radiant flux penetrating the cloud is completely and immediately absorbed at a distance of only a few tens of metres. The absorption is considerably dependent upon wavelength. It is large in the region of strong absorption where the depth of penetration is only a few metres, and small in the region of weak absorption with the depth of penetration increasing to 50–10 m. The effective radiant flux differs from zero solely within the boundary layers. Inside the cloud it equals zero and here the upward and downward fluxes are practically identical with the blackbody radiation at the temperature of the corresponding level[8].

As the radiating boundary layers of a cloud are comparatively thin, this means that in the first approximation, the thermal radiation leaving through the cloud surface can be identified with the blackbody emission at the temperature of the cloud boundary. The latter fact allows the use of the above-mentioned simple boundary condition at the cloud surface, deduced on the assumption that the thin surface is a perfect black radiator; however, it has been shown that this assumption may be considered valid only for warm clouds. Analysis of measurement data on the downward atmospheric radiation temperature at the lower cloud surface in the Arctic has shown that in this case the relative emissivity of clouds is far less than unity, especially at –6 to –10°C. This appears to be accounted for by a rapid decrease in the cloud water content and the formation of ice[9].

The above-mentioned results concern the cases of status cloud with sufficiently great optical thickness. It is clear that the radiation leaving the boundaries of the upper-layer clouds can no longer be identified with blackbody radiation. This conclusion is supported by the measurement data of Kuhn[3], for example, in the case of cirrostratus clouds with an altocumulus under layer; the emissivity turned out to be 0.75. In the presence of high cirrostatus clouds, the emissivity was 0.59. The cirrus emissivity varied from 0.10 to 0.75. In the presence of high cirrostratus clouds, the emissivity varies from 0.10 to 0.75.

The clouds can be considered as condensed matter in grand canonical ensemble. If we accept this, a broad vista is available for understanding clouds. We can imagine the clouds, atmosphere and the surface of the Earth as an ensemble and treated as a grand canonical ensemble, and clouds can be imagined as a sub-system or microcanonical ensemble.

Once we treat clouds as condensed matter, it should follow Liouvill's theorem that the Gibbs distribution function is a constant.

Assuming clouds alone to be in the grand canonical ensemble, the number of particles or phase points is a variable number of particles. The clouds contain immensely large number of particles in sub-systems and therefore the Gibbs distribution function is to be generalized in the present case for a variable number of particles. The application of Gibbs distribution function can be found in many textbooks[10].

In the canonical ensemble the sub-systems could exchange energy but not particles, with the reservoir (*r*). In a grand canonical ensemble the sub-system(s) can exchange energy as well as particles, with the reservoir. The variable N is replaced by the variable μ, the chemical potential per particle. The composite system (*c*) is again represented by a micro canonical ensemble, because the total energy E_C, and the total number of particles N_C are fixed[11].

$$E_C = E_i + E_r, \tag{6}$$

$$N_C = N_i + N_r, \tag{7}$$

$$\Delta\Gamma_C(E_C, N_C, E_S, N_r) = \Delta\Gamma_S(F_S, N_S)\,\Delta\Gamma_r(E_r, N_r), \tag{8}$$

where $\Delta\Gamma$ is the volume in space Γ.

Extension in phase in Γ space, containing definite number of phasepoints does not change with time in spite of displacement and distortion. This is the principle of conservation of extension in phase.

Let P_N be the probability in the ensemble of finding the system (*s*) in a given state *i* when it contains $N_i = N$ particles and has an energy $E_i = E_N$.

Let *s* be in one definite state *i*. Then the number of states accessible to the reservoir (*r*),

$$P_N(E_{Ni}, N) \propto \Gamma_r(E_C, N_C - N). \tag{9}$$

Since *s* is very small compared to *r*,

$$E_{Ni} \ll E_C \text{ and } N \ll N_C.$$

$$\ln \Delta\Gamma_r(E_C - E_n, N_C - N)$$
$$= \ln \Delta\Gamma_r(E_C, N_C) - \left(\frac{\delta \ln \Delta\Gamma_r(E_r - N_r)}{\delta E_r}\right)_{E_r = E_C} E_n - \left(\frac{\delta \ln \Delta\Gamma_r(E_r, N_r)}{\delta N_r}\right)_{N_r = N_C} N. \tag{10}$$

The derivatives are evaluated for $E_r = E_C$ and $N_r = N_C$ and so are constants, characterizing the reservoir (*r*). We can denote them as

$$\beta = \left(\frac{\delta \ln \Delta\Gamma_r}{\delta E_r}\right) \text{ and } \beta\mu = -\left(\frac{\delta \ln \Delta\Gamma}{\delta N_r}\right), \tag{11}$$

$$E_r = E_C, \quad N_r = N_C,$$

where the chemical potential μ represents Gibbs free energy per particle.

$$\Delta\Gamma_r(E_C - E_N, N_C - N) = \Delta\Gamma_r(E_C, N_C)\exp[-\beta(E_{Ni} - \mu N)]. \tag{12}$$

Since $\Delta\Gamma_r(E_C, N_C)$ is just a constant independent of *i* and N, eq. (9) can be written as

$$P_{Ni}(E_{Ni}, N) = C\exp[-\beta(E_{Ni} - N\mu)], \tag{13}$$

$$C = B\Gamma_r(E_C, N_C) \tag{14}$$

where B and C are constants, independent of *i* and N. This is called grand canonical distribution.

The constant C in eq. (13) is determined by normalization condition.

$$\sum_{\text{Ni}} P_{\text{Ni}}(E_{\text{Ni}}, N) = \sum_{\text{Ni}} \exp[-\beta(E_{\text{Ni}} - N\mu)]C = 1.$$

Then

$$P_{\text{Ni}}(E_{\text{Ni}}, N) = \frac{\exp[-\beta(E_{\text{Ni}} - N\mu)]}{Z}, \quad (15)$$

where $Z = \sum_{\text{Ni}} \exp[E_{\text{Ni}} - N\mu]$. Z is called the grand partition function. It is the sum of the canonical partition functions $Z(N)$ for an ensemble for different Ns, with weighting factors exp $(\beta N\mu)$.

$$Z = \sum_{N=0}^{\infty} Z(N)\exp(\beta N\mu),$$

$$Z(N) = \sum_{i} \exp(-\beta E_{\text{Ni}}). \quad (16)$$

Let us consider a grand canonical ensemble of M element ($M \to \infty$). The state of each element is characterized by the energy E_{N}, and the number N of particles in it. The statistical weight of the ensemble associated with a particular macro state $\{m_{\text{Ni}}\}$ is

$$\Omega_{\text{gM}}\{m_{\text{Ni}}\} = \frac{M!}{\underset{N}{\pi}\,\underset{i}{\pi}\, mN_i!}. \quad (17)$$

To find the most probable macro state $\{m_{\text{Ni}}\}$, we maximize $\Omega_{\text{gM}}\{m_{\text{Ni}}\}$ subject to the constraints which are generalizations of eqs (6) and (7).

$$\sum_{\text{Ni}} mN_i = M,$$

$$\sum_{\text{Ni}} mN_i E = E_{\text{C}}, \quad (18)$$

$$\sum_{\text{Ni}} nN_i = N_{\text{C}}.$$

where N_{C} is the total number of particles in the ensemble. The result is

$$\frac{\bar{m}N_i}{M} = P_{\text{Ni}} = \exp[-\beta(E_{\text{Ni}} - N\mu)] / \sum_{\text{Ni}} \exp[-\beta(E_{\text{Ni}} - N\mu)],$$

as expected with the identification of $\beta = 1/KT$ and $\alpha = -\mu/KT$ for the Lagrangian multipliers. Here α (or μ) is determined by the last condition, i.e. eq. (17). The identification of β follows from the fact that we get back the canonical distribution if we assume N to have a fixed value.

We can again define entropy by

$$S = -k \sum_{\text{Ni}} P_{\text{Ni}} \ln P_{\text{Ni}}. \quad (19)$$

(Not to be confused with the solar constant S mentioned earlier in the communication.)

Substituting eq. (15) in eq. (19) and noting that E_{N} is a function of V alone, we get

$$S = k\beta\bar{E} + k\ln Z - k\beta\bar{N}\mu$$

$$\mathrm{d}S = k\beta\mathrm{d}\bar{E} - k\beta\left[\sum_{\text{Ni}} P_{\text{Ni}}\left(\frac{\delta E_{Ni}}{\delta V}\right)\right]\mathrm{d}V - k\beta\mu\mathrm{d}\bar{N}, \quad (20)$$

$$\beta = \frac{1}{kT}, P = -\sum_{\text{Ni}} P_{\text{Ni}}\left(\frac{\delta E_{\text{Ni}}}{\delta V}\right). \quad (21)$$

We can now rewrite eq. (19) as

$$\Omega_{\text{g}} = U - T_{\text{S}} - \mu\bar{N} = -P(T, V, \mu)V$$

$$= -kT\ln Z(T, V, \mu). \quad (22)$$

$\bar{E} = U$ and Ω_{g} is the grand canonical potential which determines the entire thermodynamics. In particular

$$F \equiv U - TS = \Omega_{\text{g}} + \mu\bar{N}, \quad (23)$$

$$G \equiv U - TS + PV = \mu\bar{N},$$

$$S = -\left(\frac{\delta\Omega_{\text{g}}}{\delta T}\right)_{v,\mu}, \ \bar{N} = -\left(\frac{\delta\Omega_{\text{g}}}{\delta\mu}\right)V, T.$$

Dropping the suffixes from eqs (15) and (22),

$$P(N) = \exp[(\Omega_{\text{g}} - E + N\mu)/kT], \quad (24)$$

$$Z = \exp(-\Omega_{\text{g}}/kT)$$

$$= \sum_{N} \exp(N\mu/kT) \int \exp[-E(N)/kT]\mathrm{d}\Gamma(N). \quad (25)$$

If a proper partition function is chosen it will wholly define the behaviour of clouds to thermodynamic parameters. We quote them as follows[12]

Entropy $S = kN \log Z + \frac{3}{2} Nk,\ kN \log z + \frac{3}{2} Nk,$

Helmholtz

free energy $F = -kT \log Z,$

Total energy $E = NkT^2 \left[\frac{\delta}{\delta T} (\log z) \right]_V,$

Enthalpy $H = NkT^2 \left[\frac{\delta}{\delta t} (\log Z) \right]_V + RT,$

Gibbs potential $G = RT - NkT \log z$

Pressure $P = NkT \left[\frac{\delta}{\delta v} (\log z) + T^2 \right]_V,$

Specific heat

$$C_V = Nk \left[2T \frac{\delta}{\delta T} (\log Z) + T^2 \left(\frac{\delta^2 \log Z}{\delta T^2} \right) \right]_V.$$

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Improved content-based classification and retrieval of images using support vector machine

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Content-based image retrieval (CBIR) entails probing for similar images for a query image in an image database and returning the most relevant images. The proposed methodology aims at improving the classification and retrieval accuracy of images. Wavelet histograms are used to design a simple and efficient CBIR system with good performance and without using any intensive image-processing feature extraction technique. The unique indexed colour histogram and wavelet decomposition-based horizontal, vertical and diagonal image attributes serve as the main features for the retrieval system. Support vector machine is used for classification and thereby to improve retrieval accuracy of the system. The performance of the proposed content-based image classification and retrieval system is evaluated with the standard SIMPLIcity dataset. Precision is used as a metric to measure the performance of the system. The system is validated with holdout and *k*-fold cross-validation techniques. The proposed system performs better than SIMPLIcity and all the other compared methods.

Keywords: Colour image representation, discrete wavelet decomposition, image classification, image feature extraction.

CONTENT-based image retrieval (CBIR) finds application in a number of areas like video surveillance, medicine and geographic information system (GIS). All these applications require a high degree of accuracy with minimal user involvement. A myriad of CBIR engines have been proposed in the literature. Though most of the methods perform significantly well, the semantic gap remains to be bridged. Most of the popular methods involve region-based techniques which are computationally intensive and the success of the methods depends on the segmentation techniques used. Many relevance-based techniques have also been proposed, but the retrieved results may depend on individual perception of relevance. This spawns the need for a simple and efficient retrieval system with no user involvement.

There are several methods being used for the retrieval of images based on visual features such as colour, texture and shape. Most of the successful methods use sophisticated, time-consuming image processing techniques to

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